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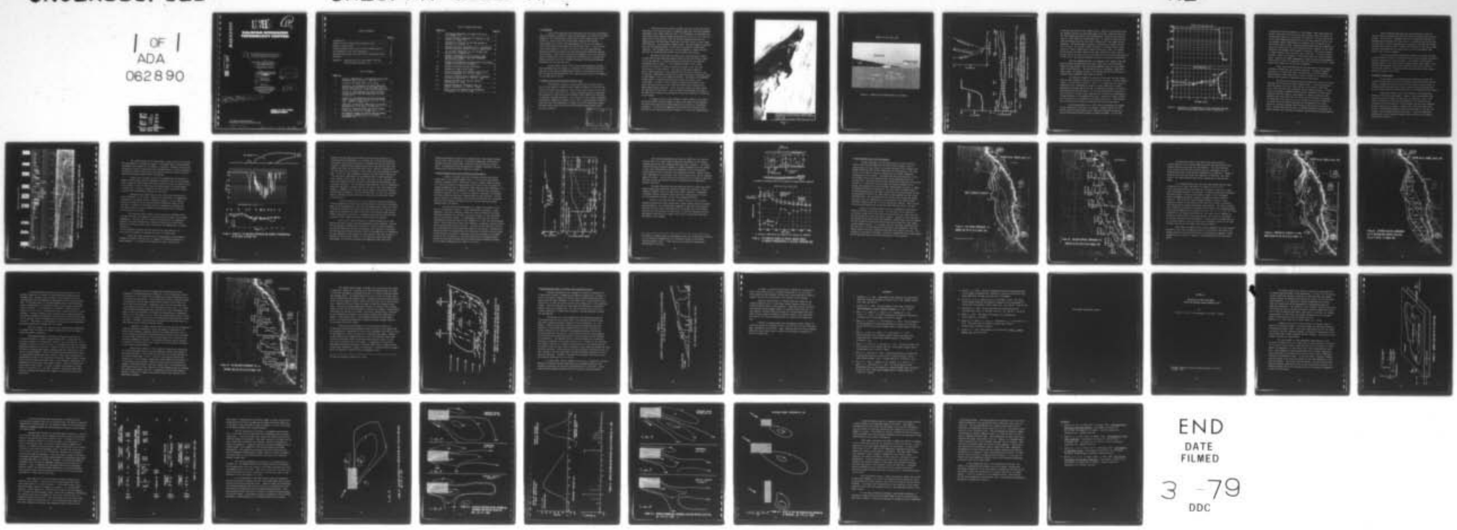

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CALSPAN ADVANCED TECHNOLOGY CENTER

(6) OFF-SHORE FOG-STRATUS SYSTEMS ALONG THE CALIFORNIA COAST.

by

(10) R.J. Pilie, E.J. Mack, C.W. Rogers, U. Katz, [redacted] W.C. Kocmond

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• Introduction

Since 1972 Calspan has been cooperating with the Naval Postgraduate School at Monterey, California, under the sponsorship of the Naval Air Systems Command, in the investigation of fog-stratus systems along the west coast of the U.S. Seven cruises have been staged aboard the NPS Research Vessel ACANIA in the waters from approximately 100 km north of Arcata to south of San Diego and extending as far as 500 km to sea. A series of contract reports (Mack et al, 1973; 1974; 1975; 1977) have been published describing specific portions of the investigation and information acquired on each cruise. These reports describe the equipment used in the investigations to provide data on the microphysical characteristics of fog and the adjacent clear air as well as micrometeorological and mesoscale phenomena associated with the fog life cycle.

In this report we will discuss the micrometeorological and mesoscale phenomena only and attempt to consolidate the information obtained to date into a descriptive, phenomenological model of the fog-stratus system. We will begin by examining briefly the conditions which lead to the formation of a new fog-stratus system.

• Formation of the Unstable Marine Boundary Layer

The prevalent situation in the eastern Pacific consists of the Pacific high typically centered several hundred kilometers to the west and south of California and producing surface winds along the coast of the order of 4 to 8 m sec^{-1} from $330^{o}T \pm 10^{o}$. Typical soundings along the coast show the existence of a well mixed marine boundary layer topped by an inversion at from 100 to perhaps 600 meters. The inversion is very strong, frequently consisting of temperature changes of up to $10^{o}C$ in approximately 100 meters. Above that level the atmosphere is normally dry with a near neutral lapse. The fog-stratus system exists within the unstable marine boundary layer below the inversion.

Destruction of the unstable boundary layer occurs when the Pacific high moves to the north and east to stimulate Santa Anna winds and strengthen subsidence in the region along the coast. The combination of subsidence and Santa Annas causes surface warming and sweeps dry air out to sea to clear the region of all cloud and fog. When the Pacific high moves back into its normal position and Santa Annas break down, the winds return to their normal 330^{o} direction. As described by Liepper (1948), the warm surface air blowing over the cold upwelling water is then cooled to produce a surface based inversion. The literature does not discuss mechanisms to explain how the inversion is raised from the surface when the Santa Annas desist.

Observations show that the fog-stratus system frequently begins to form south of San Diego and gradually moves up the coast during a three to five day interval. A typical situation is depicted by the satellite photograph shown in Figure 1. The northernmost edge of this system consists of a series of cloud or fog patches aligned in the direction of the wind as in cloud streets. The cloud or fog patches appear to grow to larger dimensions in the downwind direction until reaching a zone of clearing. Subsequently another line of almost continuous cloudiness occurs in what appears to be a well organized pattern. According to the hypothesis advanced above, a surface based inversion should exist north of this "cloud front." We have, on two occasions, maneuvered the ACANIA to the clear zone north and west of the cloud front and on both occasions have found surface based inversions. On all occasions within the cloud system, the inversion was based aloft. Here we examine the phenomena occurring at the edge of the system and the mechanism by which the inversion is lifted off of the surface.

In August 1972 we cruised beneath a stratus overcast into a patch of fog and turned the ACANIA upwind to examine the region of fog formation. This particular fog patch was wedge shaped in the vertical as illustrated in Figure 2. The low level temperature distribution and visibility data obtained in this fog are presented in Figure 3. In the clear region upwind

SKYLAB II PHOTOGRAPH OF A FOG-STRATUS SYSTEM OFF THE CALIFORNIA COAST, 1306 PDT, 2 JUNE 1973.
(Photo courtesy of J. Kaltenbach, NASA Johnson Space Center.)

Figure 1

ACANIA Sea Fog Cruise 1972

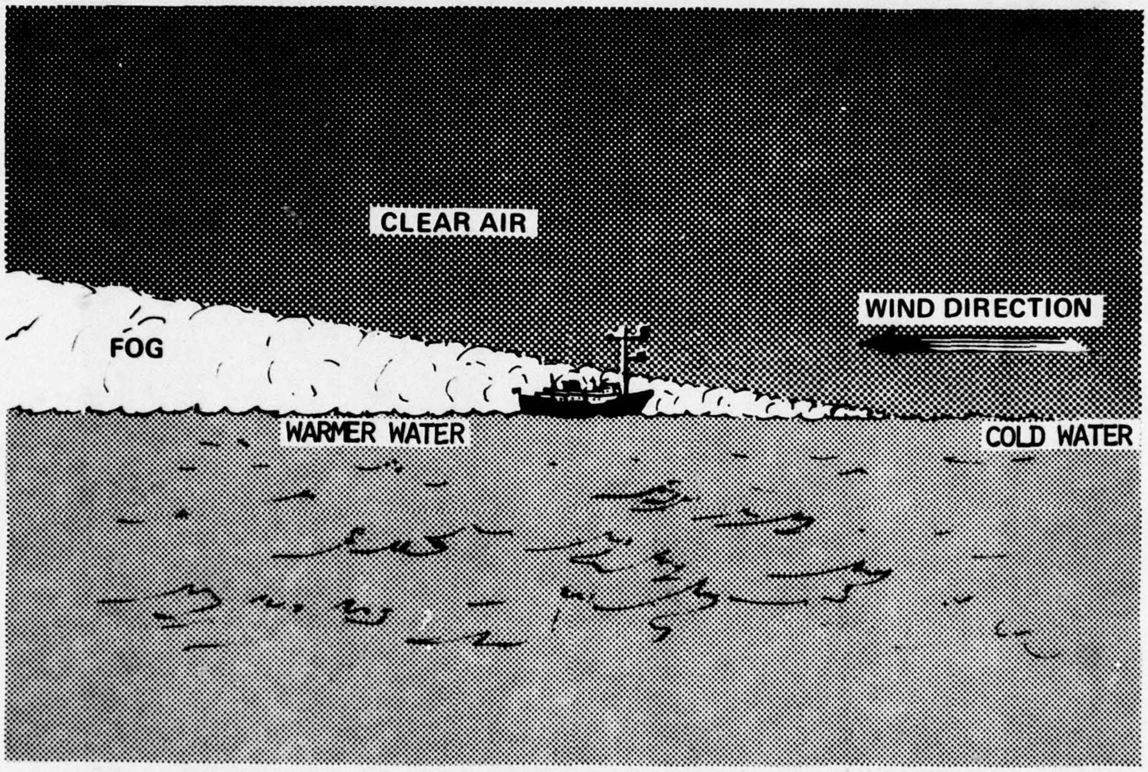

Figure 2: PROFILE OF THE UPWIND EDGE OF A FOG PATCH

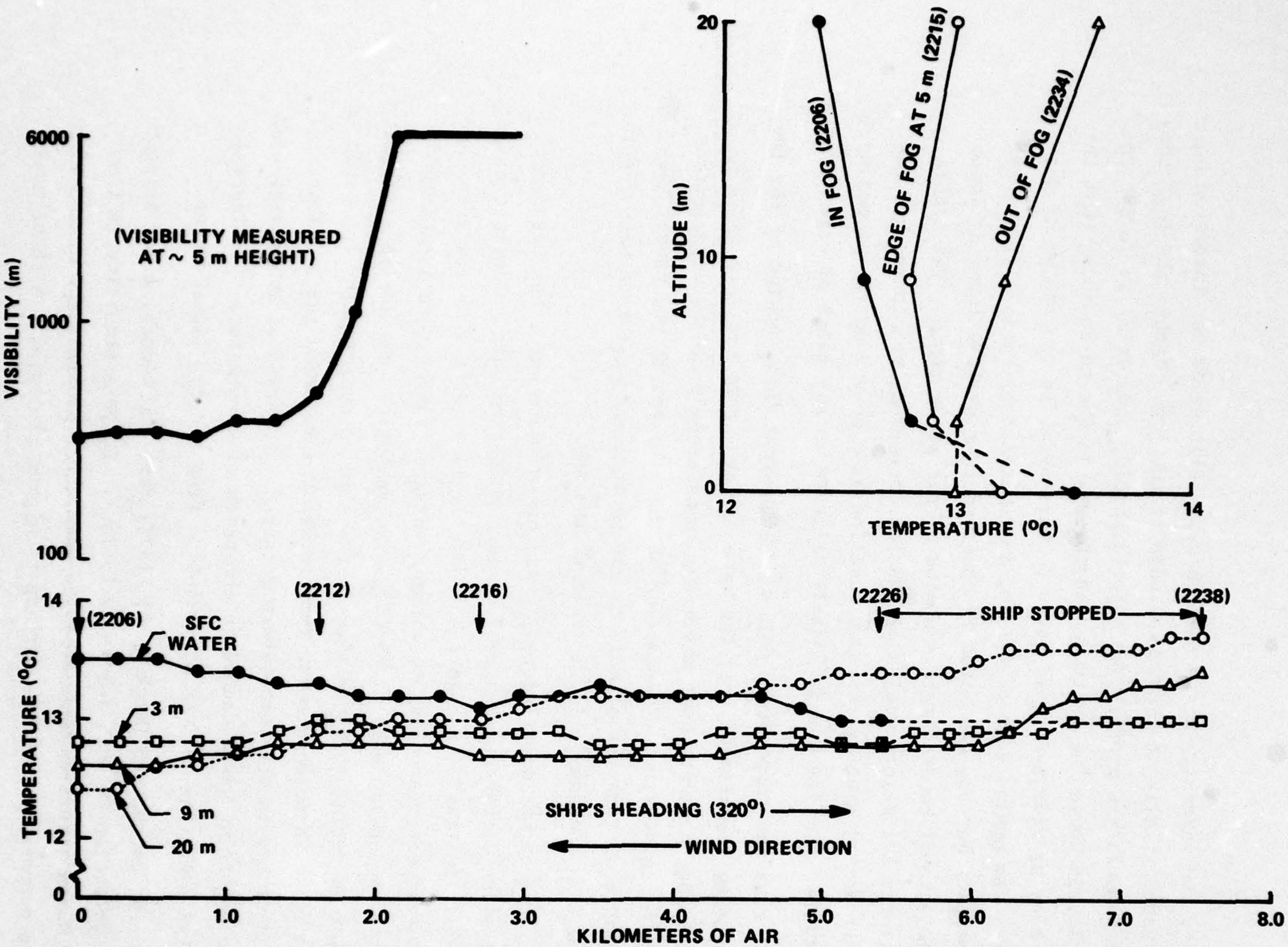

Figure 3: VISIBILITY, AIR TEMPERATURE AND WATER TEMPERATURE OBTAINED WHILE CRUISING OUT OF THE UPWIND EDGE OF A SEA FOG, 30 AUGUST 1972. INSERT SHOWS SELECTED VERTICAL PROFILES AT INDICATED POSITIONS RELATIVE TO EDGE OF FOG.

ACANIA Sea Fog Cruise 1972

of the fog, the inversion was surface based. Lifting of the inversion base was observed initially at the upwind edge of the fog. Within the fog, the low level temperature profile was superadiabatic, indicating an inversion aloft at levels above the instrumented tower. It should be noted that the wind was blowing from over cold water to warm water in the region of fog formation. The upwind edge of the fog is indicated by visibility data at the 5 m height, but fog was actually observed in a shallow layer at lower levels for several hundred meters upwind of the point shown. The ACANIA sailed several kilometers upwind of the fog, reversed its course, and re-entered the fog at exactly the same location as observed on the preceding leg of the cruise. The fog was definitely attached to the patch of warm water.

From the temperature data it is apparent that, upwind of the fog, heat was being transferred from the warm, clear air to the cold surface. Within the fog, however, heat was being transferred from the warm surface water to the colder air. The cooling of the air, therefore, cannot have been a result of transfer of heat to the surface and must have been due to radiation from condensed liquid water to the sky.

Fog formation triggered by patches of warm water has been encountered on seven different occasions on our cruises to date. The most dramatic example is illustrated in Figure 4. Here it is evident that low level air temperatures were responding to changes in water temperature as ACANIA approached the upwind edge of the fog. No measurable changes in visibility were observed during this period. As the ship encountered the fog patch, a sharp increase in surface water temperature was recorded and simultaneously a sharp decrease in air temperature occurred. Reversal of course revealed that the fog front was attached to the patch of warm water. The inversion was lifted above the surface by radiation from the fog liquid water.

We postulated (Mack et al, 1973) that the mechanism for triggering the formation of such fog is as follows. Extreme stability in the surface based inversion upwind of the warm water prevents the mixing of moisture evaporated from the surface to higher altitudes. A thin, near

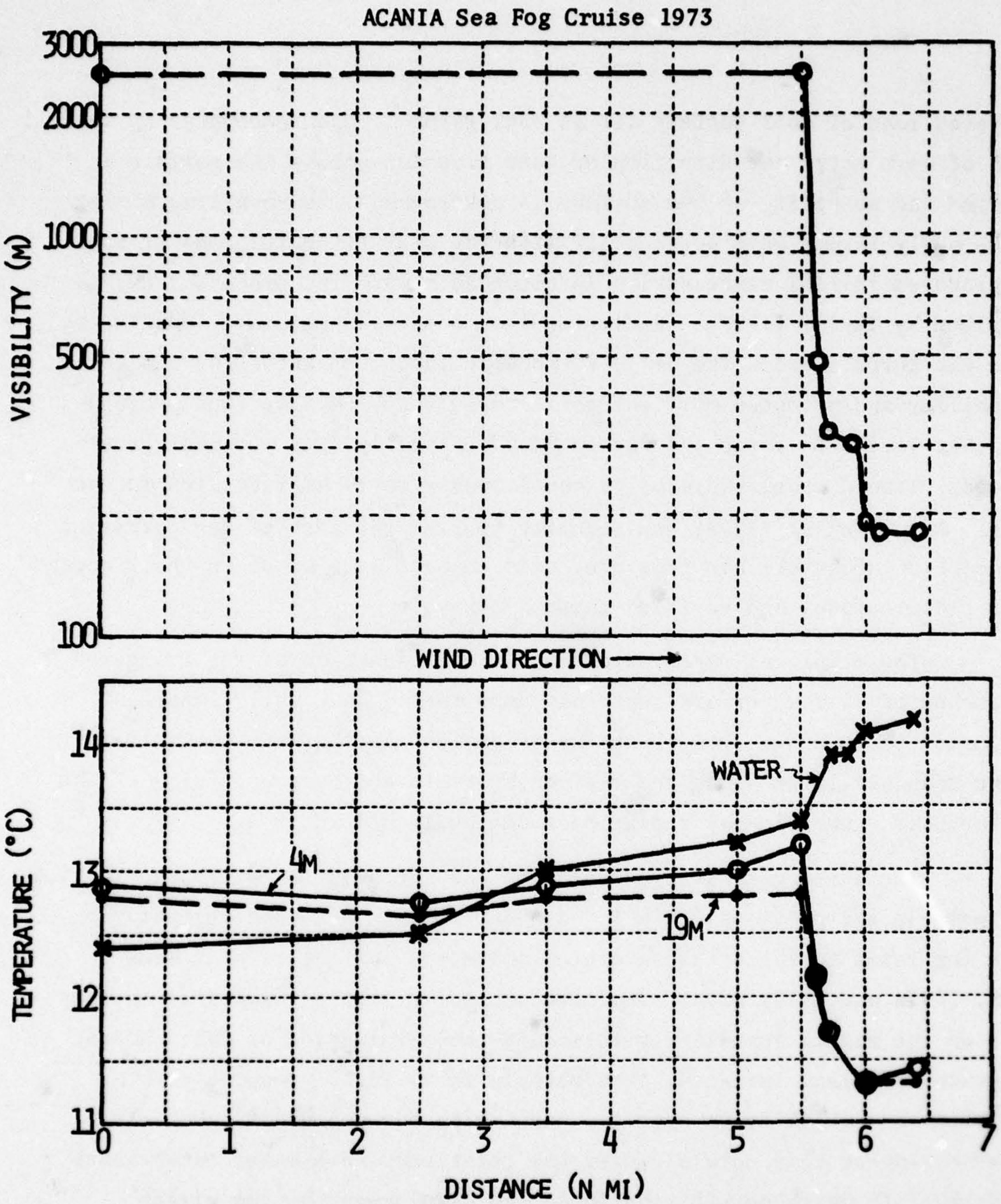

Figure 4: VISIBILITY, AIR TEMPERATURE AND WATER TEMPERATURE OBTAINED CRUISING INTO THE UPWIND EDGE OF A FOG, 10 JULY 1973.

saturated zone of cool surface air is thus formed. Upon encountering the patch of warm water the direction of heat transfer across the surface is reversed and stability at the surface is destroyed. The resulting mixing of the newly formed warm surface air with the near saturated zone of cool air produced initial condensation in accordance with the process first described by Taylor (1917). Radiation from condensed moisture immediately above the surface cools the newly formed fog layer, enhances the low level instability and promotes more energetic turbulence. Mixing thus proceeds to higher levels to force the fog to grow vertically. As vertical growth proceeds, liquid water radiates to continuously move the inversion further aloft. Oliver et al (1978) subsequently treated the case of fog formation by advection of stable air from over cold to over warm water in their second-order closure model and verified these hypotheses.

These initial observations and the explanation of fog triggered by patches of warm water are certainly important. An equally important conclusion to be drawn from these investigations is that the unstable marine boundary layer along the west coast is formed by the lifting of the surface based inversion by radiation from fog liquid water.

The above case studies describe one mechanism by which the initial fog forms in the previously stable boundary layer. A second mechanism by which fog forms to raise the inversion above the surface is what Mack et al, (1974 and 1975) have termed the coastal radiation process. In many cases on the nights immediately following the termination of Santa Annas, radiation from land surfaces, particularly in valleys, produces coastal radiation fogs which drain over the ocean with the nocturnal land breeze. The advection of this cold air over the relatively warm water establishes an instability which permits continued fog development by the mixing process as radiation proceeds from the fog top and raises the inversion aloft. Several miles off the coast, the foggy drainage air encounters the marine boundary layer wind which stimulates further mixing and fog formation.

Coastal radiation fogs have also been observed over water on numerous occasions several days after the unstable boundary layer was established. In such cases a secondary inversion is apparently formed by the drainage winds which carry the fog to sea. As radiation cools the fog top this inversion is lifted to merge with the inversion that caps the unstable boundary layer, Norton (1978). We have observed fogs of this type to form initially in Monterey Bay and either grow or drift with boundary layer winds as far as 100 km downwind.

A third process which undoubtedly produces the initial fog patch on some occasions is direct cooling from below as near saturated air advects from over warm water to over cold water. This process is described by numerous authors, e.g., Taylor (1917), Liepper (1948), Rhode (1961). While we have never observed this process along the west coast, we have one verified case study of a "cold water fog" which occurred off the coast of Nova Scotia (Mack, et al, 1976).

- Dynamics of Fog Streets

To this point we have described only mechanisms by which the initial fog forms and causes initial lifting of the inversion base off the surface. Numerous processes occur within the unstable boundary layer to stimulate production and growth of more extensive stratus cloud and fog. One of the more important of these processes is the formation of fog streets.

Cloud (fog) streets, such as are evident at the northwestern boundary of the cloud system shown in Figure 1, were encountered on two occasions during the ACANIA Sea Fog cruise of 1974 (Mack, et al, 1975). Figure 5 presents the visibility and temperature data obtained while cruising cross wind through one of these regions and shows that condensed moisture existed at the surface, verifying that these were really fog streets. Individual fog patches ranged in width from 0.5 to 2 km, consistent with the example presented in Figure 1. Vertical temperature profiles show that an unstable boundary layer existed over the region represented by Figure 5.

ACANIA Sea Fog Cruise 1974

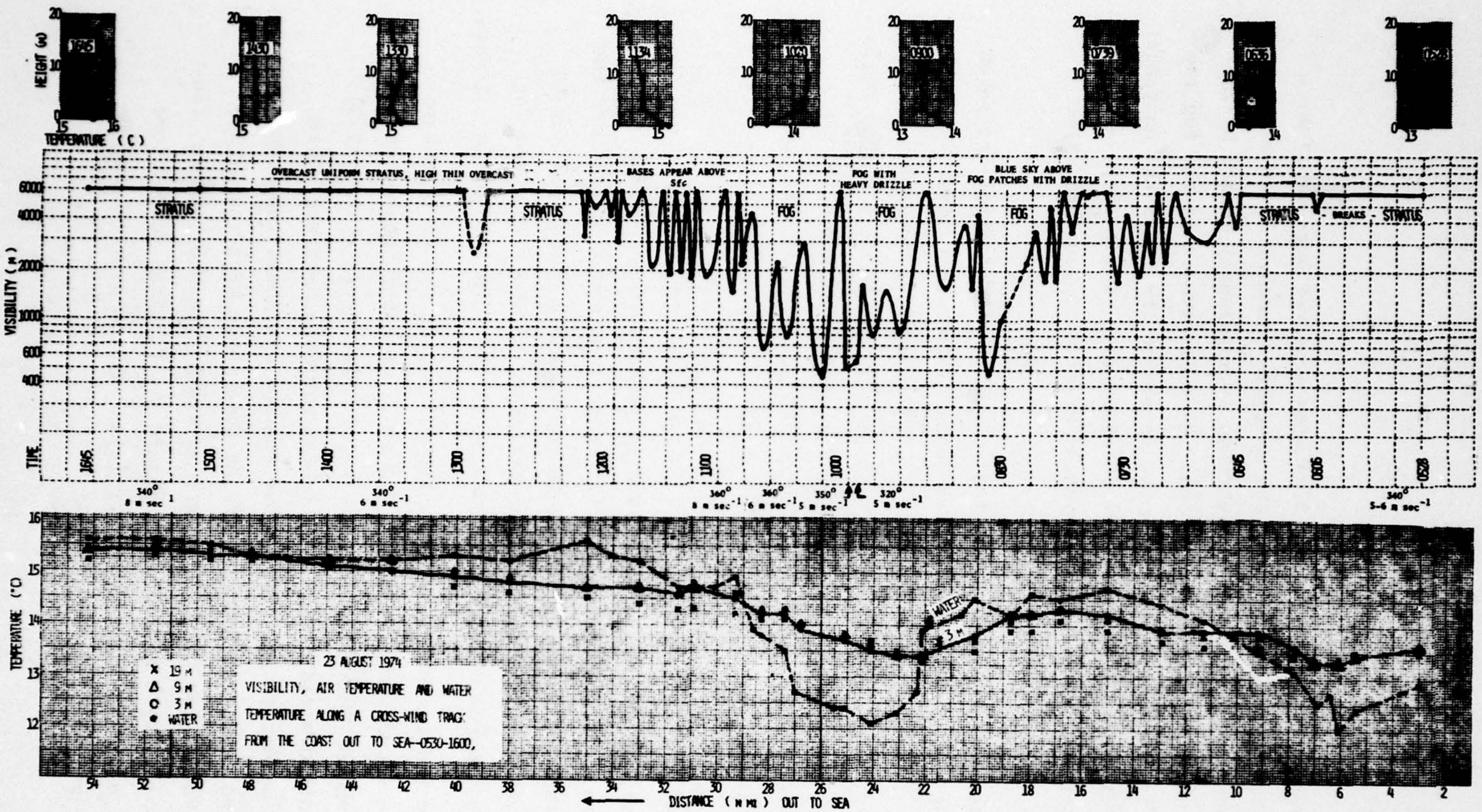

FIGURE 5 VISIBILITY, AIR TEMPERATURE AND WATER TEMPERATURE ALONG A CROSSWIND TRACK FROM THE COAST OUT TO SEA--0530 - 1600PDT, 23 AUGUST 1974

The visible profile of all fog patches, in the direction of the wind, was approximately as sketched in Figure 6. The upwind edges of the individual fog patches were attached to the surface as indicated. Downwind from the leading edge at distances ranging from several hundred meters to several kilometers, the fog base lifted off the surface and cloud persisted as stratus aloft for equivalent distances downwind before dissipating.

Figure 6 also includes visibility and temperature data obtained as ACANIA cruised upwind along the approximate centerline of one of these patches. After leaving the upwind edge of the fog we cruised for some ten nautical miles in an attempt to enter the next fog patch and succeeded in cruising beneath a stratus tail. However, a change in sea state forced a reversal of course to protect the instrumentation mounted on ship's bow.

Temperature data showed that a truly surface-based inversion was never encountered in this region. Upwind of the fog patch, air temperatures responded to very gradual surface temperature changes. We were unable to explain the formation of the shallow fog patch immediately upwind of the main patch. Formation of the larger patch was apparently associated with the sharp increase in water temperature occurring around the 14 nautical mile position on the figure.

Two important points should be noted from these temperature data.
(1) Within the fog, air temperature did not respond to changes in sea surface temperature. Instead, from the 16 nautical mile position, air temperature gradually increased downwind over a region of substantially colder water. Surface air warming was more closely correlated with improvements in surface visibility.
(2) The atmosphere downwind from the fog patch was approximately 0.5C cooler than the surface air entering the fog at its upwind edge.

While these data constitute our only measurements cruising upwind through the patches of fog streets, we suggest the following processes as being important. As described in the previous section, the clear air

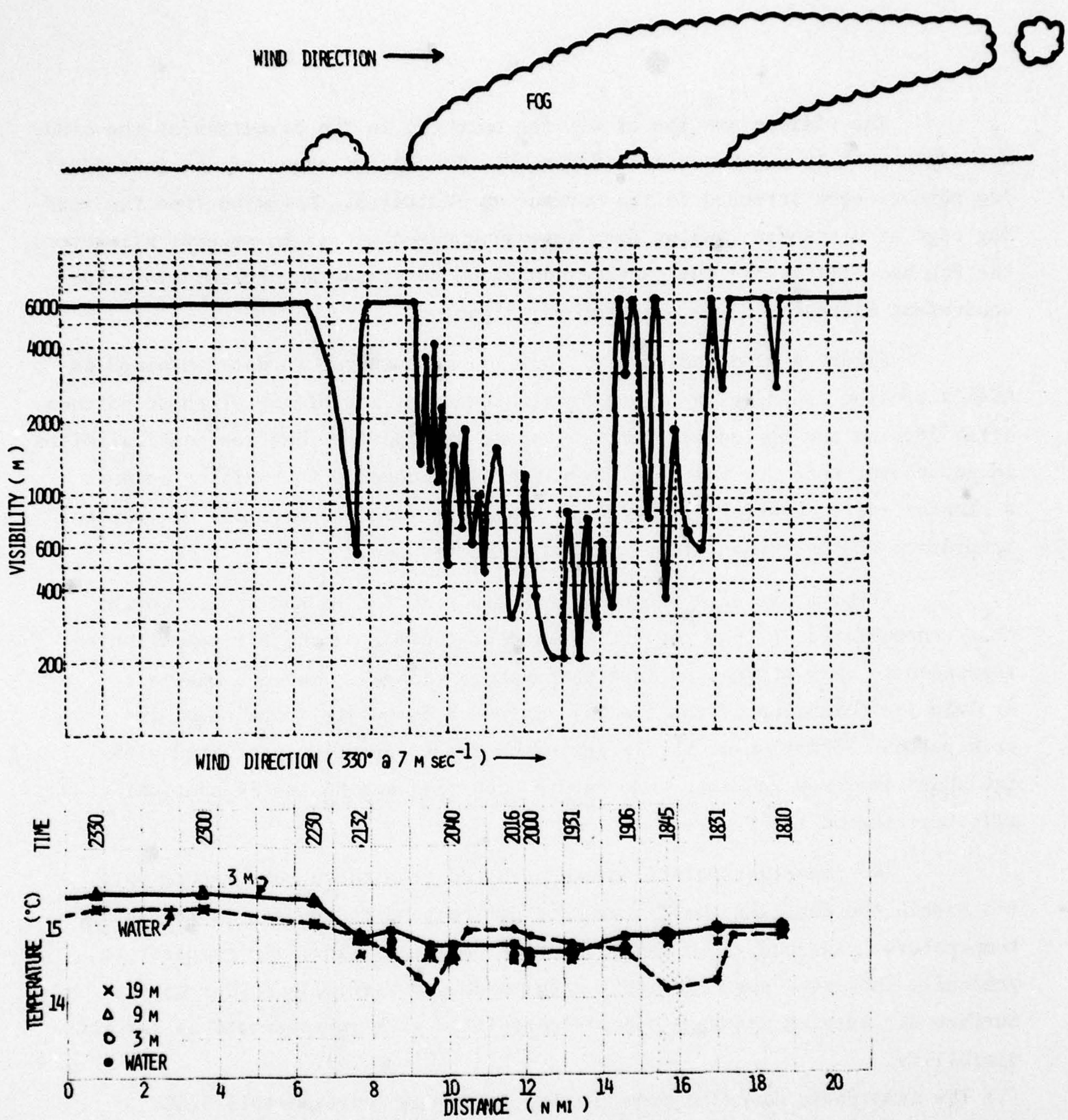

FIGURE 6: VISIBILITY, AIR AND WATER TEMPERATURE AND SCHEMATIC REPRESENTATION OF FOG PATCH, 24 AUGUST 1974

upwind of the fog stratus system advects over cold water where it is conditioned for fog formation, i.e., surface air is cooled and moistened. An encounter with a patch of warm water stimulates mixing and initial fog formation. Radiation from condensed water raises the inversion above the surface only where the initial fog patch exists. With the dissipation of this fog patch, the air downwind is left cooler than that entering its upwind edge. As the low level atmosphere encounters the next patch of warm water, it is now better conditioned for fog formation than air flowing between initial patches, and a larger fog forms. This fog, in turn, dissipates but leaves the boundary layer air even more conditioned for fog formation; and the process continues. The likelihood of fog formation in the air immediately behind the fog patch is greater than on each side so that cloud streets are produced. Gradually, the individual patches become so large that adjacent patches merge and produce a band of continuous fog. By this time, radiation from the continuous fog top maintains and enhances the continuous, stable inversion aloft.

In private discussions, Dr. James Telford of the Desert Research Institute suggested that the reason for dissipation of individual fog patches might be the development of a very localized high pressure region as a result of radiative cooling of the well mixed region within the fog. Advection at upper levels maintains the stratus aloft, while the localized high pressure inhibits advection at low levels. Considering the localized nature of such a pressure field, significant dynamic effects should be expected.

Oliver et al, (1978) attribute the daytime evaporation of lower portions of stratus clouds to direct absorption of solar radiation. Because of the difference in absorption coefficient of clouds at long wavelengths (terrestrial radiation) and short wavelengths (solar radiation), they point out that long wave radiation from the cloud top overcompensates for absorption of solar radiation in that region and causes stratus cloud tops to persist. The bases of thick clouds, however, are in radiative balance with their surroundings at terrestrial radiation wavelengths,but short wave solar

radiation penetrates to that level to produce direct cloud droplet warming and low level cloud dissipation. This process undoubtedly contributes to dissipation of the trailing edge of the fog, leaving persistent stratus aloft. A combination of the two processes is probably responsible for the characteristic fog patch profile depicted in Figure 6.

- <u>Stratus Formation Aloft and the Stratus Lowering Process</u>

A variety of dynamic processes determine where fog and stratus clouds form in the unstable boundary layer. In the earliest publication available, Anderson (1931), showed that California coastal stratus occurred as a result of adiabatic cooling whenever the inversion base rose above the lifting condensation level. Petterssen (1938) using the same techniques, i.e., temperature and humidity soundings by aircraft, verified these conclusions. Anderson recognized, that as a result of radiation, instability is developed just below cloud tops and causes downward growth of the cloud base. If the initial inversion is sufficiently low and radiation persists for a sufficient period of time, Anderson concluded that this "stratus lowering" process produced fog at the surface.

We have observed these phenomena on a number of occasions at sea. One case illustrated in Figure 7 vividly demonstrates the process. Under completely cloud-free conditions, ACANIA was cruising westward (cross wind) from Point Conception and recorded a continual increase in inversion height with the NPS acoustic sounder. At approximately 0735 local time, initial condensation was observed aloft in what appeared to be very thin fracto-cumulus clouds. Within minutes the broken cover merged to a completely overcast stratus deck stretching from horizon to horizon. By 0800, initial visibility restrictions were observed as the stratus base lowered to the surface. Fog and solid stratus persisted for approximately five hours until 1300 when the first break in the overcast was observed. Subsequent calculations showed, as the figure indicates, that the presence of the fog-stratus layer corresponded almost exactly with the period during which the inversion base exceeded the height of the lifting condensation level (Mack et al, 1977).

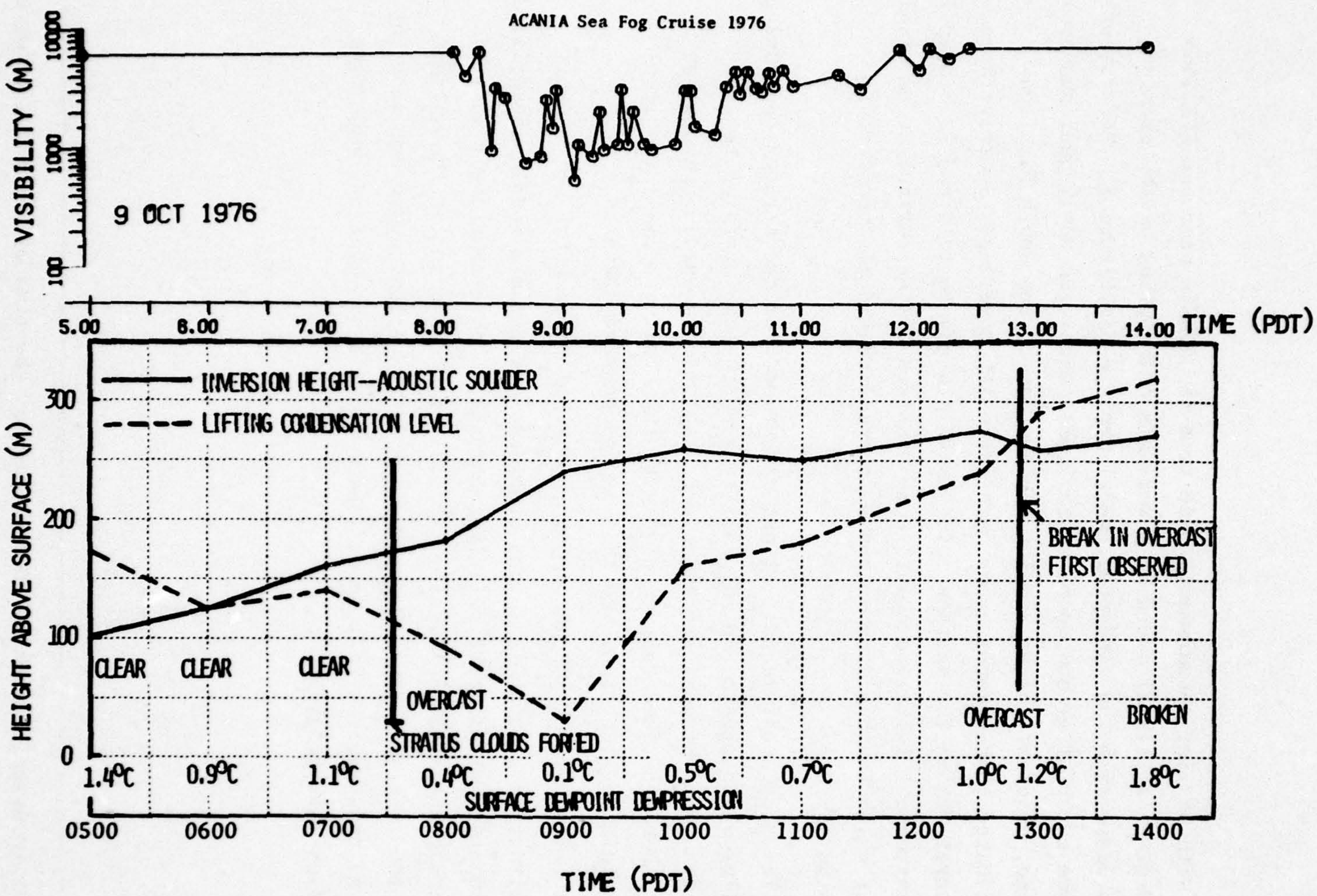

FIGURE 7 VISIBILITY, INVERSION HEIGHT, AND LIFTING CONDENSATION LEVEL FOR FOG OF 9 OCTOBER 1976

With earlier measurements made from the ACANIA near the Farallon Islands, Mack et al (1973) added to Anderson's description of the stratus lowering process with the conceptual model depicted in Figure 8. Persistent stratus from early afternoon appeared to remain unchanged until near sundown. At that time, observations of the 100 m high peak of the south Farallon Island indicated that stratus bases were gradually lowering. Shortly thereafter temperatures in the surface layer to 20 m began to fall and 20 m dewpoint began to increase. Initial droplets were observed at the surface at precisely the time at which the humidity increase* terminated and air temperature decrease stabilized.

The model postulated that net radiation from the stratus top caused rapid cooling which created an instability beneath the inversion to cause turbulent transport of cool air and cloud droplets downward. Evaporation of droplets beneath cloud base coupled with the cooling lowered the level at which saturation occurred, and the base propagated downward.

Oliver et al (1978) performed a detailed analysis of the stratus lowering situation with their second order closure dynamic model and verified these postulates. It is particularly important to note that their analysis indicates that the stratus lowering process can produce fog at the surface only when the inversion base is below 350 m, in excellent agreement with the experimental observations of Leipper (1948) and all observations made from the ACANIA on this program.

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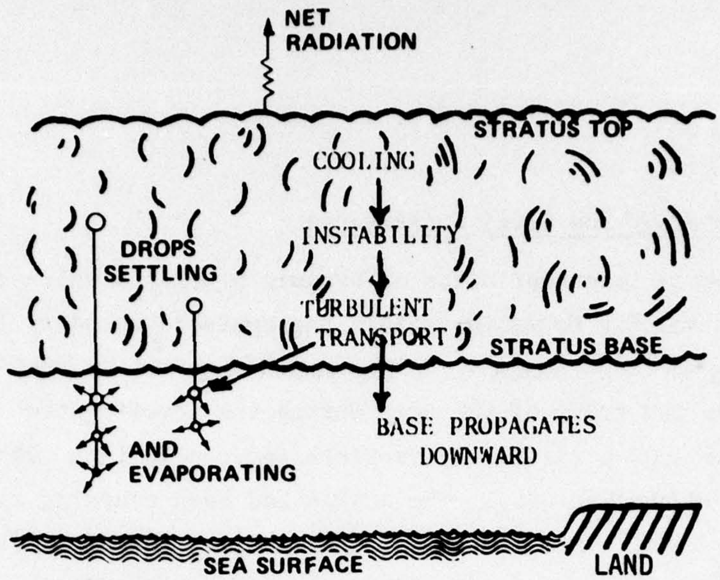
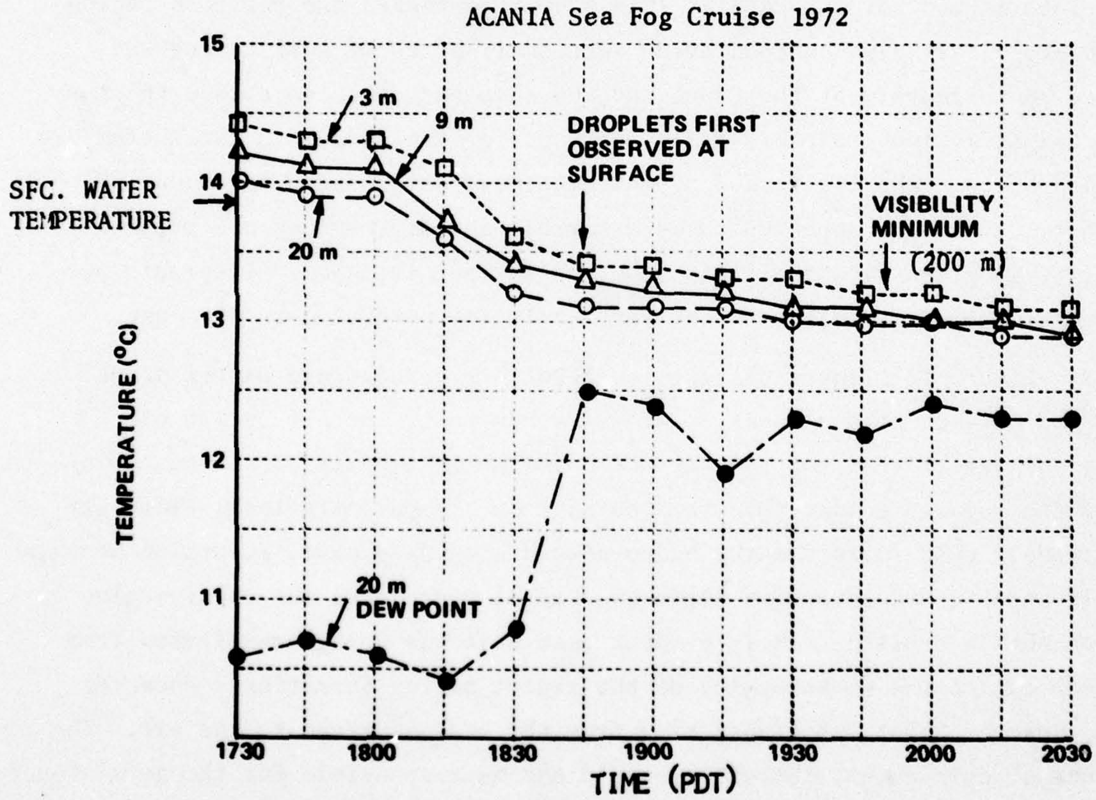
*The offset in dewpoint from air temperature at the time of dense fog occurrence at the surface was probably due to a calibration offset of the lithium chloride dewpoint sensors. These devices do not function properly when contaminated by sea salt.



a) SCHEMATIC REPRESENTATION OF FOG FORMATION THROUGH STRATUS LOWERING



b) LOW-LEVEL TEMPERATURE AND DEWPOINT DATA DURING FOG FORMATION

FIGURE 8: FOG FORMATION THROUGH THE STRATUS LOWERING PROCESS
A) SCHEMATIC REPRESENTATION; B) LOW-LEVEL TEMPERATURE DATA

• <u>Fog Stimulated by Low Level Convergence</u>

Another important class of dynamic processes which stimulates stratus cloud and fog formation within the unstable boundary layer is illustrated by the case study of a fog which occurred on 26-27 August 1974. Figure 9 shows the track of the ship during the investigation of this fog as well as the 1000 m visibility isopleth and sea surface temperature distribution in the vicinity. The ACANIA had been cruising within a 150 x 150 kilometer area to the north and west of Cape Mendocino during the previous week. Fog and stratus had been encountered through most of that week, so that it is certain that we were within the unstable boundary layer. During late morning of the 26th as ACANIA cruised toward the coast, a region of complete clearing was encountered. One small patch of surface fog was observed approximately at the Cape, and the ship was directed toward the fog. At approximately 1600 PDT this small patch of fog exploded into the system described below. The fog formed over the coldest water encountered during the cruise, and there appeared to be a correlation of the thousand meter visibility isopleth with the 12 to 13^{o}C sea surface isotherm. Clear air persisted over the warmer water to the west while fog persisted to the east.

Figure 10 illustrates the low level air temperature distribution inside and outside of the fog system. In every case, the air upwind of the fog was warmer than the sea surface below while the air within and downwind of the fog was colder than the sea surface. Apparently local flow over the extremely cold surface water had produced a surface based inversion beneath the well established inversion aloft and had also produced the clear region in which ACANIA cruised. It is evident that heat was being transferred from the clear air to the water upwind of the region of fog formation. Once fog formed, however, heat was transferred from the ocean surface to the air. The existence of cold water, therefore, could not be responsible for the persistence of this fog patch over the 20 hours during which it was observed.

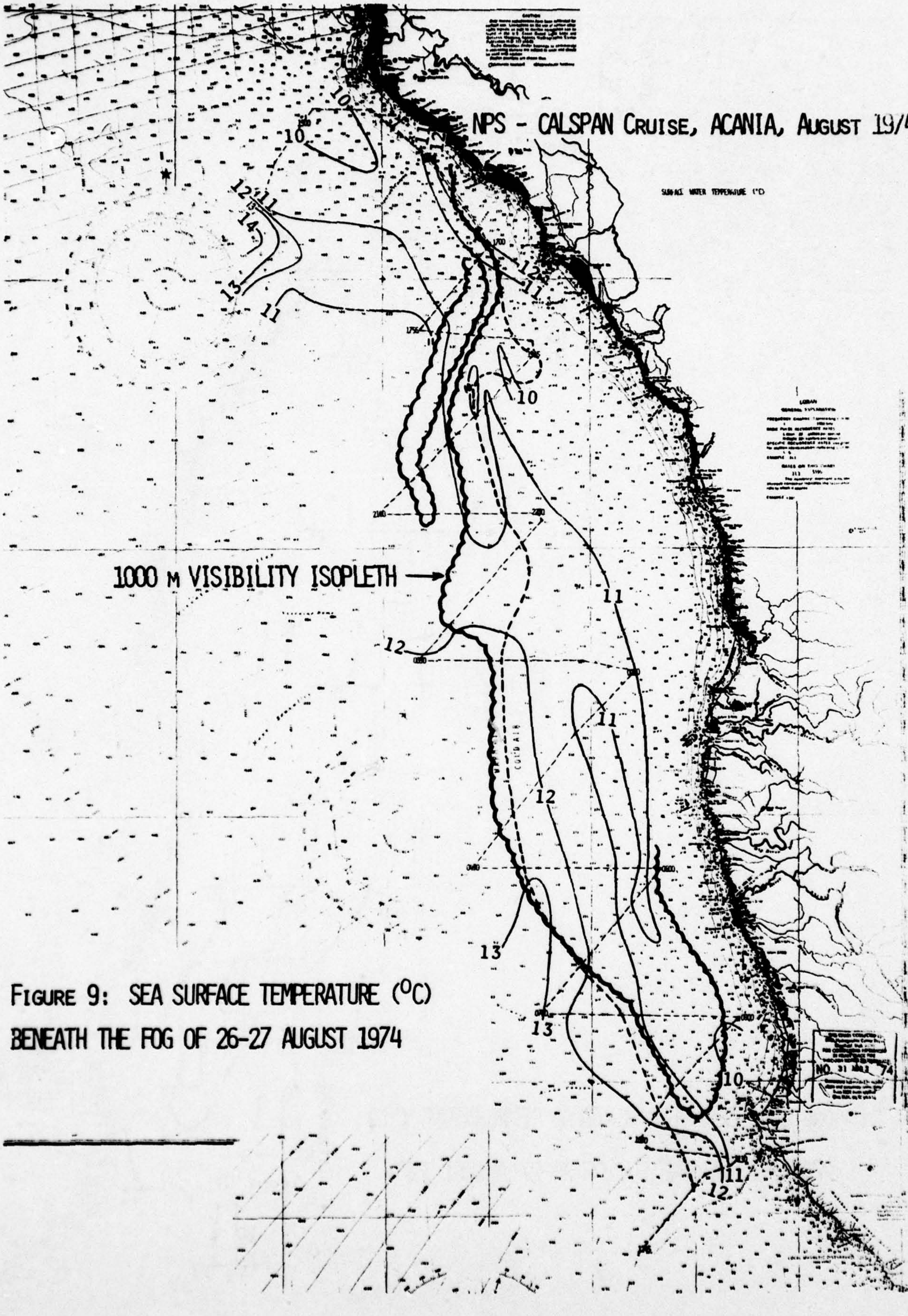


FIGURE 9: SEA SURFACE TEMPERATURE (^{O}C)
BENEATH THE FOG OF 26-27 AUGUST 1974

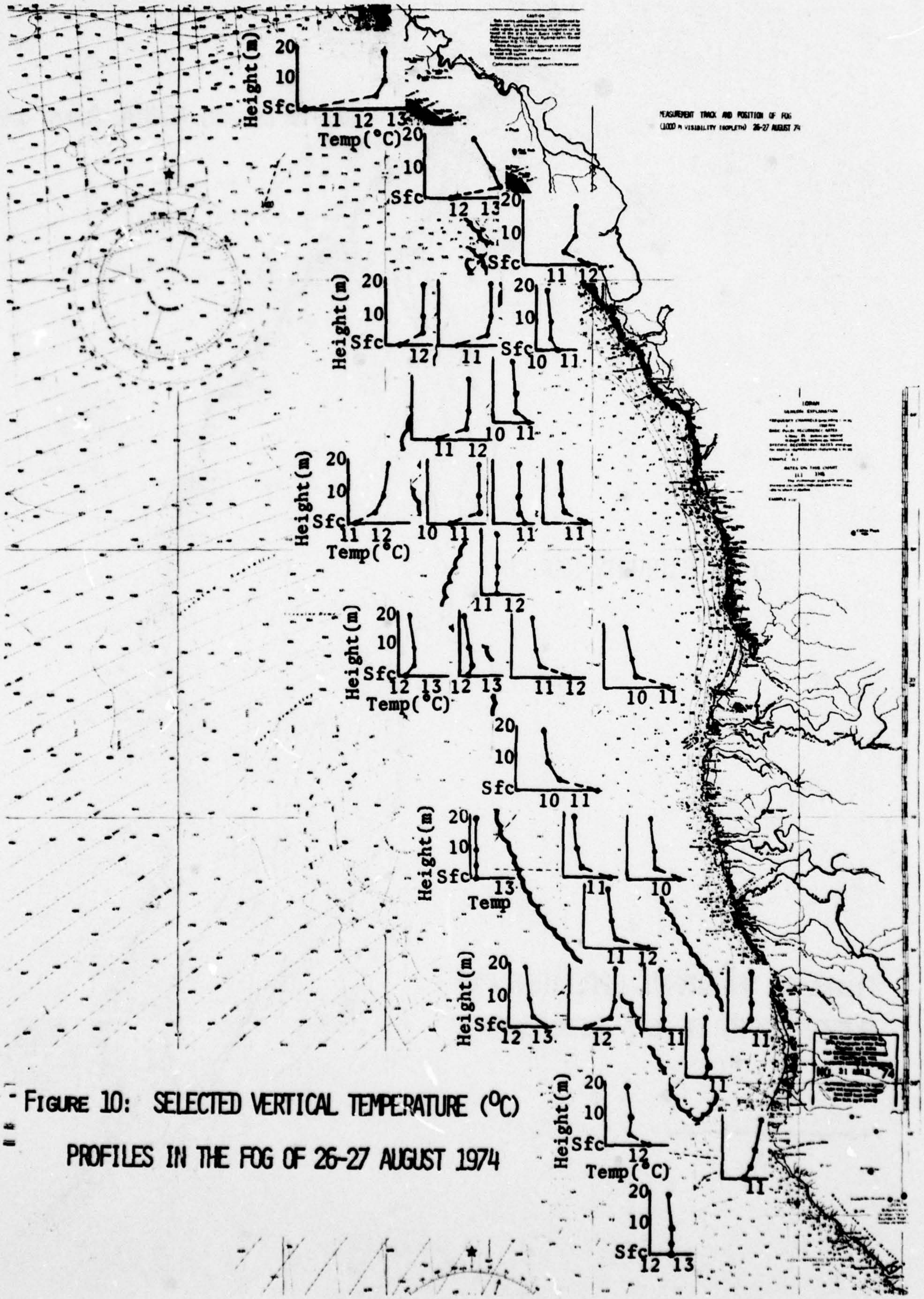


FIGURE 10: SELECTED VERTICAL TEMPERATURE (°C) PROFILES IN THE FOG OF 26-27 AUGUST 1974

The particular ship's track used in this case study was selected because it was suspected from previously obtained data that convergence had been responsible for fog formation. This track permitted us to stop the ship and obtain the best possible measurements of wind speed and direction at each location considered to be important. Figure 11 presents these wind data along with isopleths of visibility. The scale shown in the legend provides quantitative information on observed wind speeds.

It is apparent from these data that winds upwind from the region of fog were in all cases from $\sim 330^{o}$ with speeds of 4 to 8 m sec^{-1}. Within the region of fog, wind speeds were consistently lower and in some cases directional reversals were observed. A region of persistent convergence obviously existed. Using the data shown in Figure 11, the convergence pattern throughout the fog was computed by graphical means as shown in Figure 12. The verticies of each triangle sketched in Figure 12 were permitted to advect at the measured wind velocity for a short, arbitrarily selected period of time and area changes were computed. The results indicate that, except for the region of obviously chaotic winds in the northern portion of the fog, divergence values ranging from -0.7×10^{-4} to -2.7×10^{-4} sec^{-1} were determined for all regions within the fog. The actual values computed are indicative of the levels of divergence present, but are not as significant as the fact that all values were negative. To the extent that these data can be considered synoptic, persistent low-level convergence existed for at least 20 hours over the area of roughly 2500 square kilometers.

Using these convergence values and the assumption that wind velocity measured at the 20 meter height persisted to an altitude of 100 meters, the average vertical velocity within each of the triangles was computed. Values ranged from 1 to 2 cm sec^{-1}, always in the upward direction for the entire period.

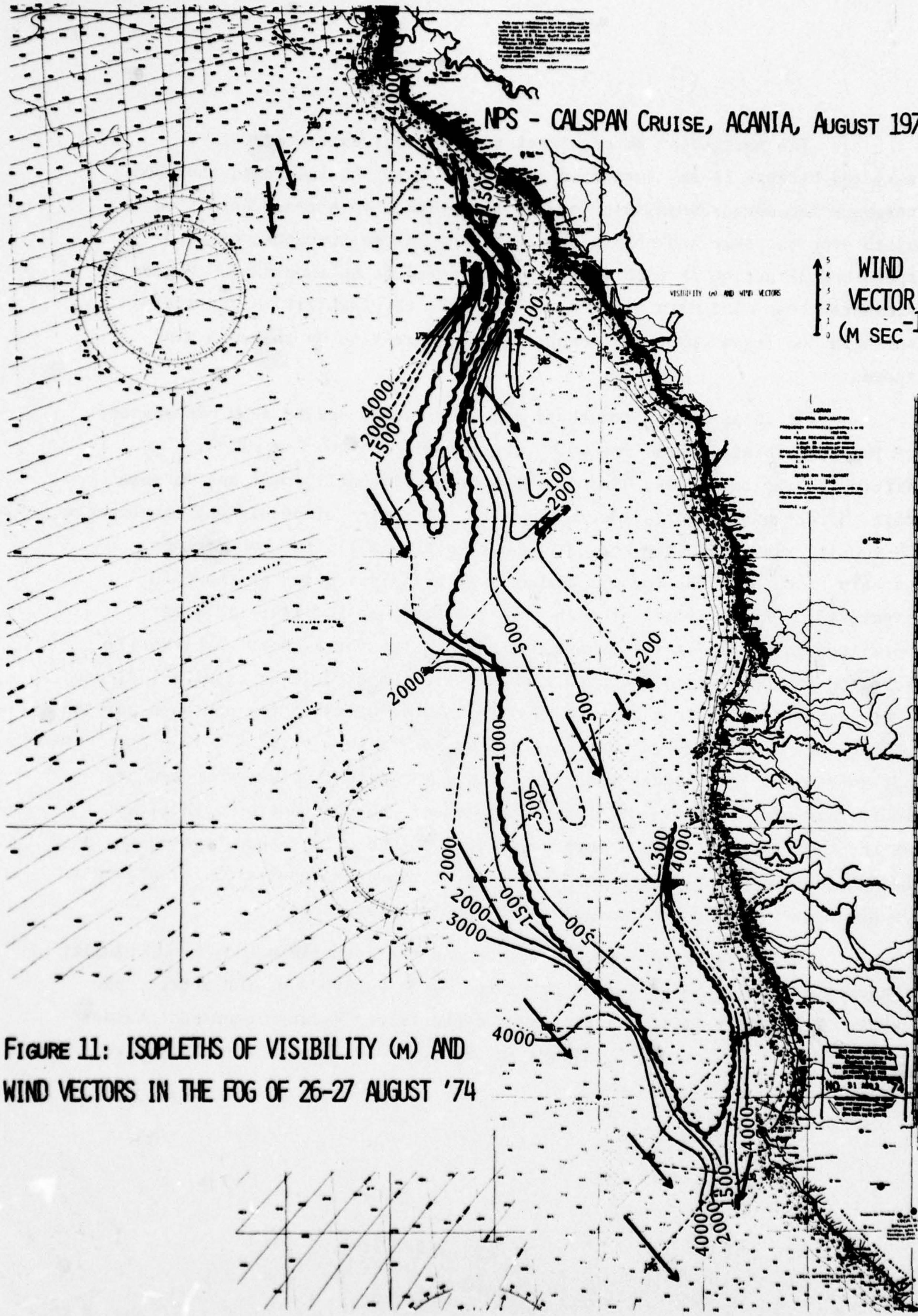


FIGURE 11: ISOPLETHS OF VISIBILITY (M) AND WIND VECTORS IN THE FOG OF 26-27 AUGUST '74

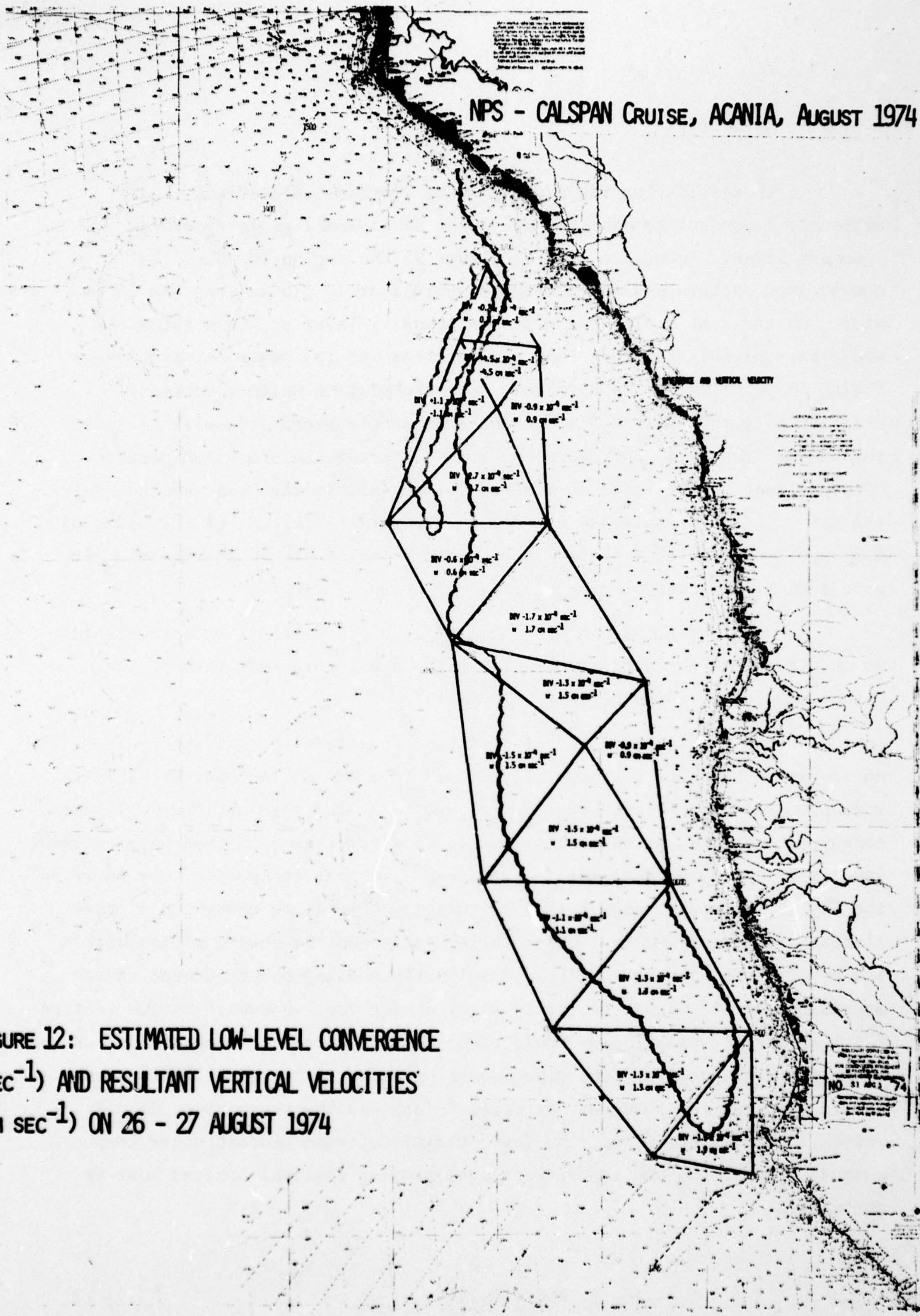


FIGURE 12: ESTIMATED LOW-LEVEL CONVERGENCE (SEC^{-1}) AND RESULTANT VERTICAL VELOCITIES ($\mathrm{CM\ SEC}^{-1}$) ON 26 - 27 AUGUST 1974

In considering the nature of the vertical velocities, it is extremely important to consider the three dimensional geometry of the boundary layer. Recognize that in Figure 9, the region in which the convergence pattern was calculated is approximately 110 km long and 24 km wide. On the same scale, the unstable boundary layer of 400 m thickness would be substantially less than the thickness of the paper on which the figure is printed. It is difficult to contemplate a uniform vertical velocity of 1 to 2 cm sec^{-1} extending over such a broad area with so little thickness. The data, therefore, must indicate that an organized pattern of persistent updrafts and downdrafts existed within the area and that only the net vertical velocity averaged over the entire area was of the order of 1 cm sec^{-1}. Individual up- and downdrafts existing within the region must have been significantly stronger than the average.

With the existence of a persistent net upward motion established, we inquired into the effects of adiabatic expansion in this thin region using some further assumptions.

The Oakland sounding of 26 August indicated the existence of an inversion based at a height of approximately 400 m. Assume that this structure extended to the experimental region at the same altitude. If net convergence existed to the 100 m level, net divergence must then have existed immediately beneath the inversion. Assume also that the surface air entering the fog at the upwind edge ranged in humidity from 85 to 95% which is consistent with measurements. Under these conditions the upward motion within the experimental region would cause adiabatic cooling and condensation of approximately 600 mg m^{-3} of liquid water at fog top. Downward motion (return flow) within the same region would cause, by wet adiabatic compression, an evaporation of precisely the same amount of water. The result would be no liquid water at the surface. In order to explain the presence of liquid water at the surface (i.e., the fog), therefore, some process other than adiabatic expansion and compression accompanying vertical motions must be postulated.

An important clue was derived from the measured low level temperature distributions shown in Figure 13. The air temperature within the fog was 1 to 2^{o}C colder than air temperatures upwind regardless of the wind trajectory considered. It was shown earlier that the observed decrease in temperature could not have resulted from transfer of heat from air to water within the fog region. Furthermore, heat could not have been lost in the upwind direction to the warmer atmosphere on that side. It is known from the Oakland sounding that heat could not have been transferred to the warm air above the inversion and from previous measurements at Vandenberg (Mack, et al, 1972; Rogers, et al, 1974) that during daylight hours at least, the hills to the east of the fog were substantially warmer than the air over the ocean. Therefore, the only process by which heat could have been lost from the fog was by radiation from the fog top.

Radiation, of course, produces cooling at fog top and, consequently, produces additional condensation. While the amount of liquid water condensed by wet adiabatic expansion during the ascending portion of the air flow is evaporated during the descending portion of the air flow, the amount of liquid water produced by radiative cooling remains to be carried to the surface by downward air motions. Thus radiation from the fog top explains the presence of liquid water at the surface in the same manner as described in the stratus lowering process.

Quantitative estimates of the amount of liquid water expected at the surface were made based on observed temperature changes from upwind to within the fog region. It was shown earlier that total cooling observed at the surface ranged from 1 to 2^{o}C. If the air entering the fog had been saturated, a 1 to 2^{o}C change in temperature would produce 800 to 1600 mg m^{-3} at the surface within the fog. These values far exceed the liquid water contents of 350 and 100 mg m^{-3} observed, respectively, in the northern and southern portions of the fog. However, with the assumption that air entering the northern and southern zones was at approximately 95% and 85% relative humidity, respectively, computed and observed liquid water contents are in approximate agreement.

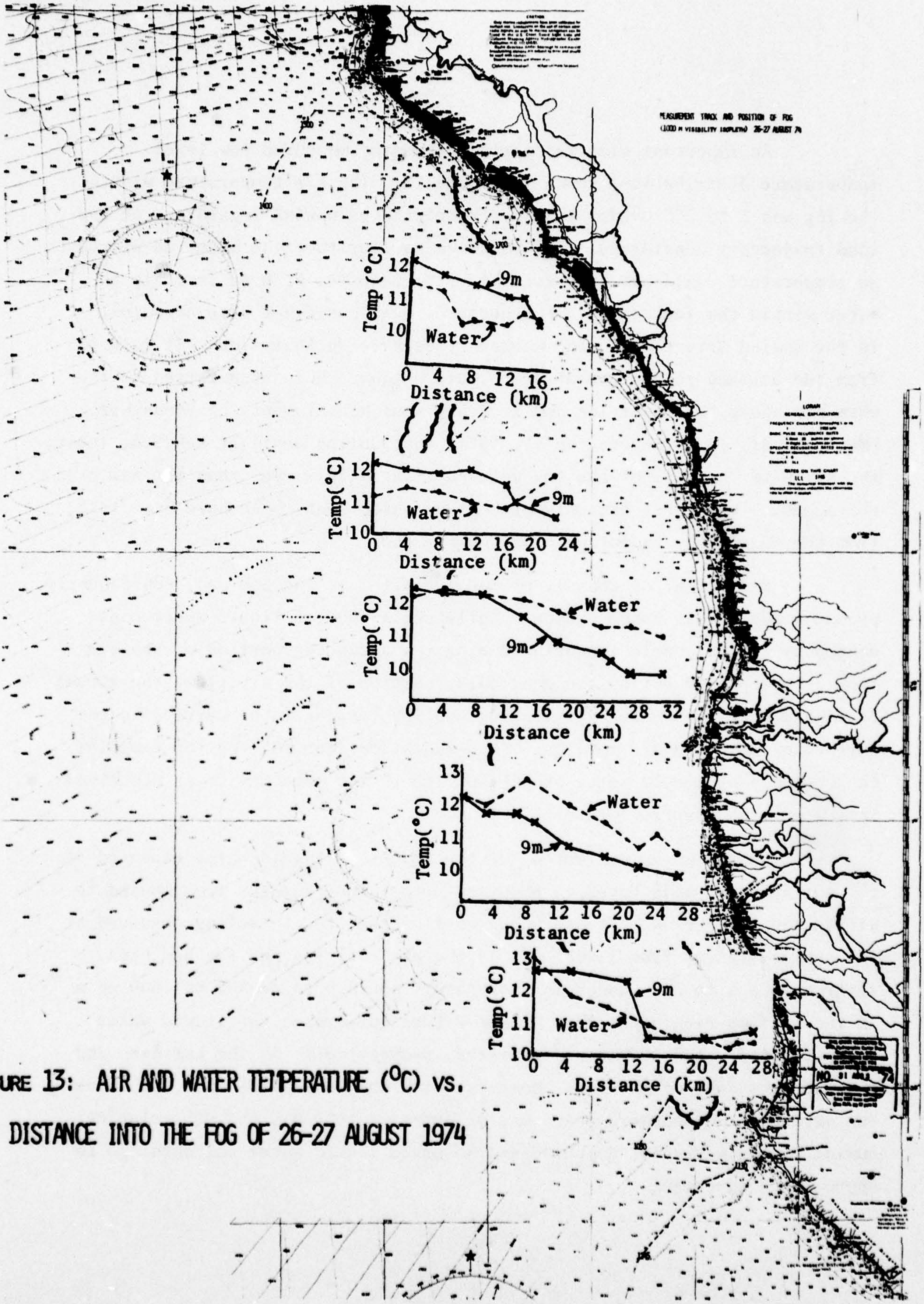


FIGURE 13: AIR AND WATER TEMPERATURE (°C) VS. DISTANCE INTO THE FOG OF 26-27 AUGUST 1974

The overall model of what is thought to be occurring in this region is depicted in Figure 14. Incoming air at near saturation enters the region of low level convergence and becomes involved in a system of organized upward and downward motion. The figure depicts only one cell of this system, which must be comprised of many adjacent cells. Expansion during the upward portion of the air flow produces stratus clouds aloft with their tops at the inversion base. Radiation from the stratus top causes cooling and increased liquid water. The extreme cooling immediately beneath the inversion base increases stability above the fog top and produces instabilities beneath. Simultaneously, liquid water at fog top is increased. The return, downward flow, stimulated by this instability, produces a wet adiabatic warming which evaporates an amount of liquid water equal to that produced during the initial upward motion. The liquid water produced by radiative cooling, however, persists to constitute fog at the surface.

From satellite photographs of this fog stratus system*, it is evident that the particular convergence pattern studied on 26-27 August 1974 was produced by an interaction of the low level atmosphere with the hills south at Cape Mendocino. Hence the land mass is included in the model. However, we are convinced that numerous processes other than frictional effects may stimulate convergence within the unstable marine boundary layer, so that land is not a necessary part of the model. In Appendix A, we describe a preliminary modeling investigation of dynamic effects from a region of warm surface water imbedded in a region of cold water, a situation which we suspect may be an important stimulus of local convergence at sea.

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*On file at Calspan; see Mack et al, 1975.

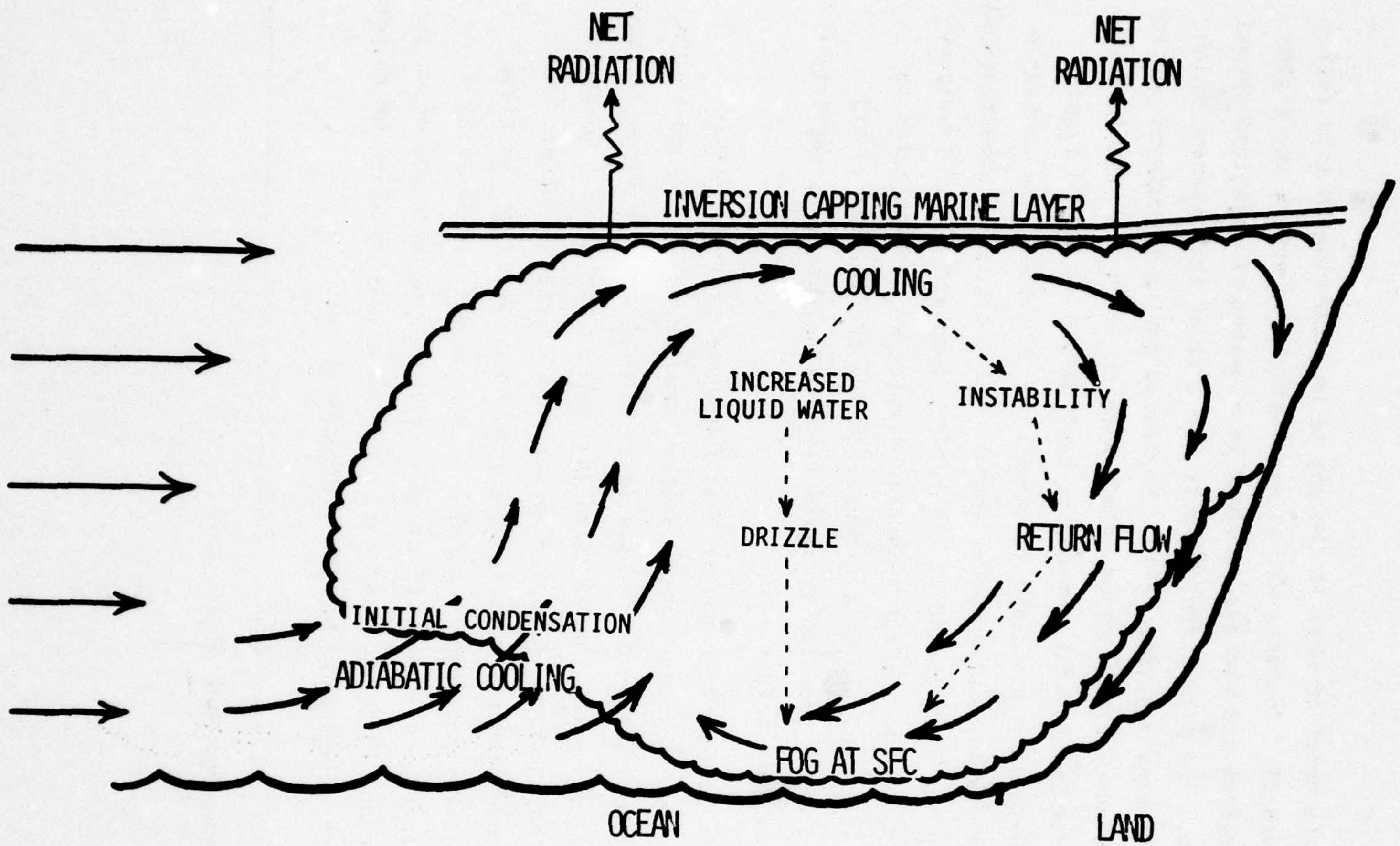


FIGURE 14: SCHEMATIC REPRESENTATION OF THE VERTICAL CROSS SECTION OF FOG FORMED BY LOW-LEVEL CONVERGENCE AND RADIATIVE COOLING

• Phenomenological Model of the West Coast Fog-Stratus System

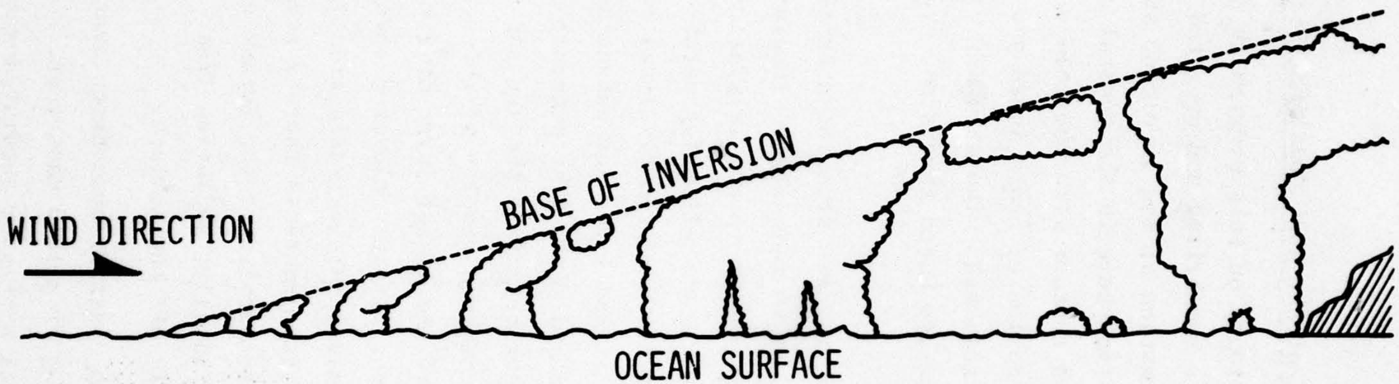
The model developed on this program to describe the organization of fog-stratus systems is summarized and depicted schematically in Figure 15. Prior to the initial formation of fog or cloud, subsidence and Santa Anna winds clear the marine atmosphere in the coastal zone. Both processes cause warm dry air to come into contact with the cold surface water in the upwelling regions. With the termination of Santa Annas and subsidence, northwesterly surface winds are re-established. The warm air blowing over the cold sea surface establishes a surface based inversion, creating conditions suitable for fog formation.

Initial fog is at times stimulated by the subsequent passage of the cooled surface air over warmer water. The instability created by the warm water stimulates mixing to produce initial condensation, i.e., fog patches. Radiation from these shallow fog patches establishes the local inversion at fog top and further promotes local low-level instabilities, thereby producing a well-mixed layer and enhancing exchange of heat and moisture between the air and sea. (The presence of the cold fog represents a sink for moisture evaporating from the sea and therefore further enhances net evaporation.) The combined results of these phenomena are accelerated cooling (by radiation) of the lowest layer of air, a transfer of the inversion base from the surface to a slightly elevated level, the accelerated addition of moisture to the air mass, and near-adiabatic lapse conditions beneath the local low-level inversion. These processes thereby accelerate conditioning of the air mass, priming it for more persistent fog formation farther downwind, so that a fog street develops. Gradually, radiation from sequential fog patches raises the inversion permanently off the surface.

At other times, initial condensate over water is produced by nocturnal drainage of radiation fog from land masses and subsequent mixing of cold, saturated continental air with near saturated marine air over water. Again, radiation from condensed moisture raises the inversion from the surface.

Figure 15:

SCHEMATIC REPRESENTATION OF ORGANIZATION OF FOG-STRATUS SYSTEMS
OFF THE COAST OF CALIFORNIA



NPS-CALSPAN CRUISES, ACANIA, 1972-1976

No doubt, on some occasions initial condensation is produced by direct cooling from below as air advects from warm to very cold water. This process has been observed off Nova Scotia and postulated by numerous authors, but has not been observed in the Pacific on this program.

Once the inversion base is elevated from the surface and the unstable boundary layer is established, numerous dynamic processes can stimulate fog and stratus cloud formation. Important processes that have been observed include gradual changes in inversion height such that, in some regions, inversion height may exceed the lifting condensation level. There, stratus clouds are formed by adiabatic cooling. When conditions are correct, the stratus lowering process can convert these clouds into fog.

Patterns of strong surface-level convergence produce regions of persistent upward and downward air flow, and, with the inversion sufficiently high, condensation occurs aloft. Again, radiation from the stratus produces additional liquid water which is carried to the surface by downward air motion to produce fog.

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APPENDIX A

APPLICATION OF THE LAVOIE MODEL WITHIN THE UNSTABLE MARINE BOUNDARY LAYER*

By

Roland J. Pilié, C. William Rogers, and James T. Hanley

*Calspan Corporation Internal Research Report, No. 85-416
December, 1978

The overall objective of the Navy's marine fog investigation is to develop improved methods of fog forecasting. In view of the conclusions in this report that observed variations in inversion height and dynamic processes within the boundary layer govern the location of fog and stratus cloud formation, it is important to identify mechanisms which control these two characteristics and approaches for incorporating them into prediction procedures. The similarity of the boundary conditions associated with marine fog stratus systems and boundary conditions employed by Lavoie et al (1970) in modeling lake effect snow storms is striking. Our pevious investigations of lake effect snow storms and application of the Lavoie model to that problem met with substantial success. (McVehil et al, 1968; Eadie et al, 1971a, Eadie et al, 1971b).

A modified version of the Lavoie model, we reasoned, might be applied to predicting changes in inversion height and convergence patterns in the marine boundary layer. Since model applications did not fall within the scope of our Navy sponsored programs, we pursued the application of the Lavoie model under Calspan sponsored internal research. The results of that internal research are reported here.

The Lavoie model is a two dimensional representation of the horizontal characteristics of a single layer air mass capped by an inversion. In its simplest form, it may be described with the aid of Figure A1, which depicts its application to lake effect snow storms originating over Lake Erie. Input parameters require specification of the surface temperature at each grid point and specification of wind velocity and air temperature entering the experimental region. In the lake effect situation, boundary layer air upwind of the warm lake surface has the same temperature as the cold land surface. The initial inversion height is specified along with a constant potential temperature of the air mass beneath the inversion, the potential temperature at the inversion base, and the potential temperature at the top of the model.

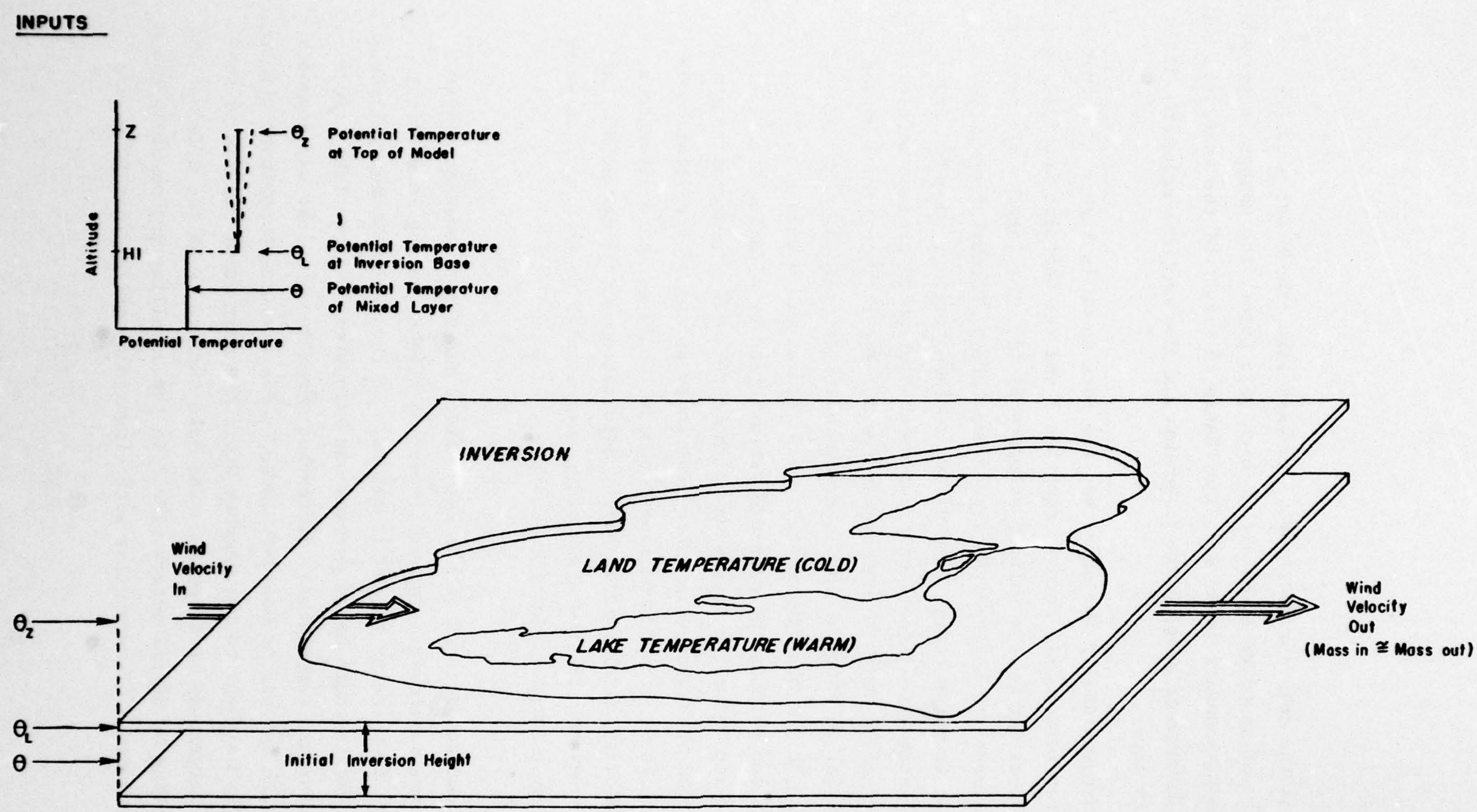


FIGURE A-1: BOUNDARY CONDITIONS FOR THE LAVOIE MODEL

As the cold air moves over the warm lake, the model predicts the velocity of boundary layer air at each grid point, the change in potential temperature of boundary layer air, the change in height of the inversion and diagnoses divergence beneath the inversion and the vertical velocity within the boundary layer.

Mathematically the model has the form shown in Figure A2. Initially the advection term equals zero in equation 1 and the Coriolis and frictional terms are specified by the initial velocity field. The large scale pressure force is selected to balance the latter two terms. Perturbation pressure forces arise from two terms. The direct surface heating, computed from equation 3, stimulates localized pressure changes which produce accelerations (equation 1) and a non-zero divergence field. The last term in equation 4 is the vertical velocity at the inversion base, determined by the product of inversion height and divergence. This term causes a change in the inversion height and extends the pressure perturbation by altering the fraction of air beneath and above the inversion in the model. The inversion height deformation term is included in subsequent computations of acceleration in equation 1. Thus, it appeared that the model could estimate divergence, vertical velocity and inversion height deformation in the marine boundary layer which may result from air blowing over a patch of warm surface water surrounded by a large field of cold water.

Some comparisons of the lake effect and marine boundary layer situations are worthwile. Typical inversion heights during lake effect storms range from one to three kilometers. In the marine boundary layer, inversion heights range up to perhaps one kilometer. In the lake effect situation, surface temperature differences from land to lake may approach 10°C. Beneath the marine boundary layer, surface water temperature differences are typically only one to two degrees C off the west coast but may be substantially larger off Nova Scotia. The lake surface is of the order of 500 kilometers long by 50 kilometers wide. In the upwelling region along the west coast, patches of warm water with dimensions smaller than 50 km are

$$\frac{\partial u}{\partial t} = \underbrace{-\vec{V}\cdot\nabla u}_{\text{ADVECTIVE TERM}} + \underbrace{f v}_{\text{CORIOLIS TERM}} \underbrace{- C_{DRAG}|\vec{V}|u}_{\text{FRICTIONAL FORCE}} \underbrace{- \frac{1}{\rho}\frac{\partial P}{\partial X}}_{\text{LARGE SCALE PRESSURE FORCE}} \tag{1}$$

$$\underbrace{\underbrace{+ \frac{g}{\theta_L}\left[\theta - \frac{\theta_L + \theta_Z}{2}\right]\frac{\partial H}{\partial X}}_{\text{INVERSION HEIGHT DEFORMATION}} + \underbrace{\frac{gH}{2\theta}\frac{\partial \theta}{\partial X}}_{\text{SURFACE HEATING}}}_{\text{PERTURBATION PRESSURE FORCES}}$$

$$\frac{\partial v}{\partial t} = \text{SIMILAR TO } \frac{\partial u}{\partial t} \tag{2}$$

$$\frac{\partial \theta}{\partial t} = \underbrace{-\vec{V}\cdot\nabla\theta}_{\text{ADVECTIVE TERM}} + \underbrace{C_{HEAT}|\vec{V}|(\theta_{SURFACE} - \theta)}_{\text{SURFACE HEATING}} \tag{3}$$

$$\frac{\partial H}{\partial t} = \underbrace{-\vec{V}\cdot\nabla H}_{\text{ADVECTIVE TERM}} \underbrace{- H\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right)}_{\text{VERTICAL VELOCITY (DIVERGENCE) TERM}} \tag{4}$$

FIGURE A-2: MATHEMATICAL FORM OF LAVOIE MODEL

more typical. Considering the consistent change in overall scale of the characteristics of the two systems and the highly successful application of the model to the lake effect storms, preliminary exercise of the model on the marine boundary layer seemed warranted.

For the initial test we selected a rectangular patch of water (40 x 15 km) which was $2^{o}C$ warmer than the surrounding water of uniform temperature. Initial wind speed was 5 meters per second and the initial inversion height was 300 meters. From these conditions, the model was run until steady state conditions were achieved. Results are presented in Figures A3 through A5. Figure A3 represents calculated deformation of inversion height, which amounted to 85 m between maximum and minimum deformations. Such variations are typical of those observed with the Naval Postgraduate School acoustic sounder aboard the ACANIA on numerous occasions; however, the observations have never been tied to specific patches of warm water.

The inversion deformation is presented in the upper portion of Figure A4. The middle portion of that figure presents the divergence field computed while the lower portion of the figure represents the field of computed vertical velocities. Both sets of values are consistent with those determined from the data acquired in a fog observed south of Cape Mendocino in 1974 (see Figure 12 in text).

It is interesting to note that the peak values of divergence and vertical velocity extend to the left of the undisturbed wind and downstream from the warm perturbing surface. Figure A5 shows the distribution of inversion deformation and vertical velocity in the downwind and cross wind directions through the point of maximum upward deformation. Note that maximum vertical velocities are located in regions of maximum change in inversion height. If fog and/or stratus are produced by these vertical velocities, we should expect it to be observed first in this region.

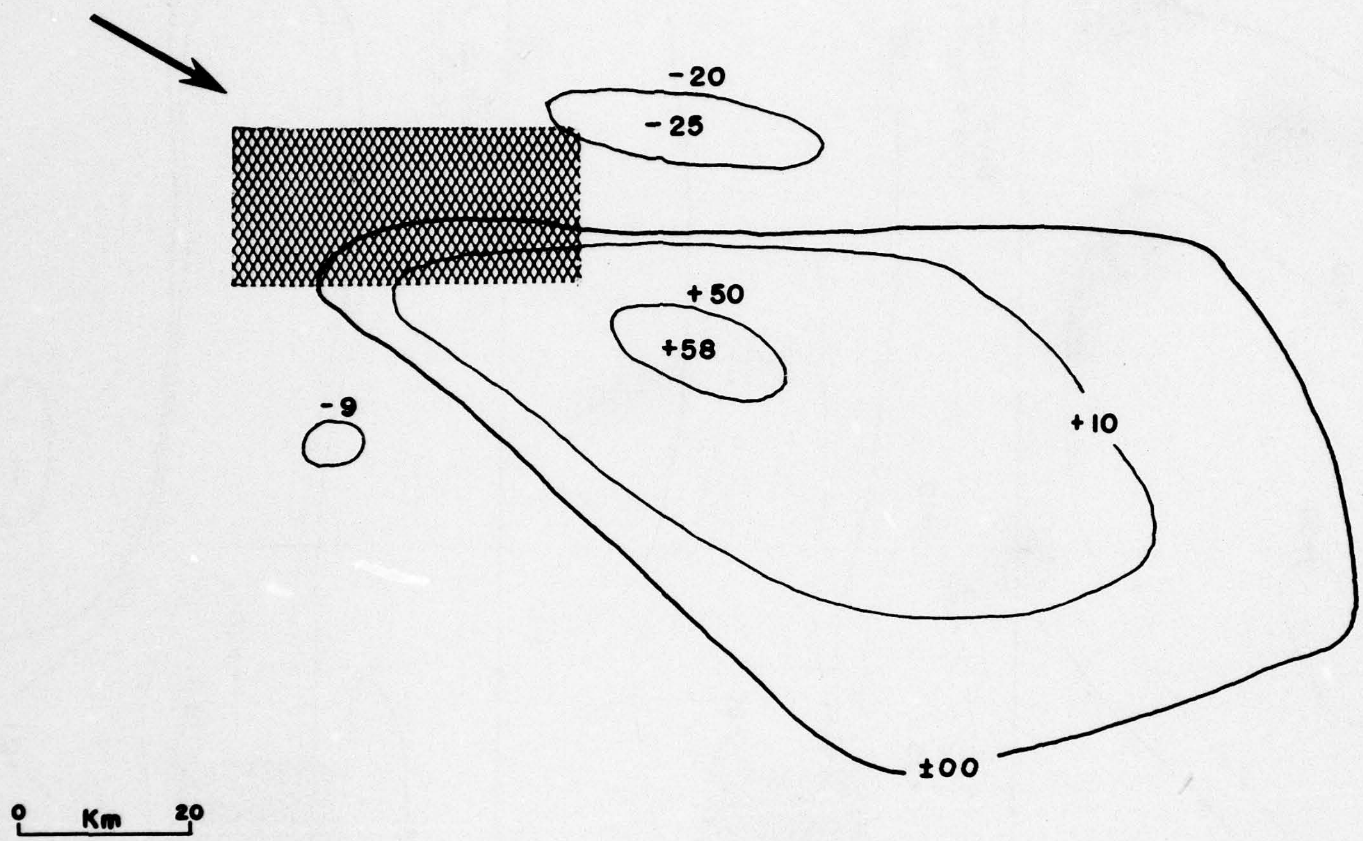


FIGURE A-3: CALCULATED DEFORMATION OF INVERSION HEIGHT DUE TO PATCH OF WARM WATER.
$\Delta T = 2^{o}C$, HI = 300M

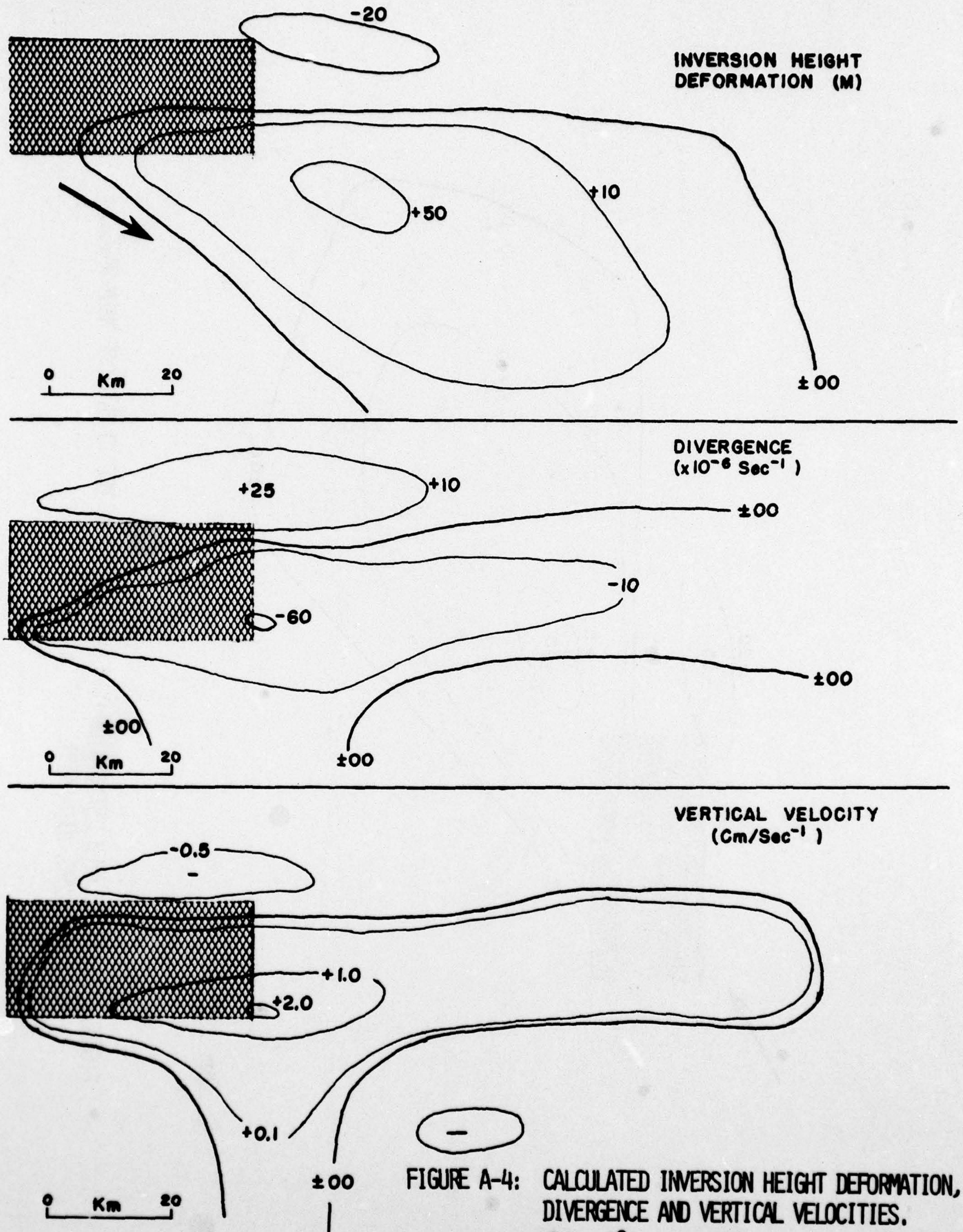


FIGURE A-4: CALCULATED INVERSION HEIGHT DEFORMATION, DIVERGENCE AND VERTICAL VELOCITIES. $\Delta T = 2^{o}C$, HI = 300M

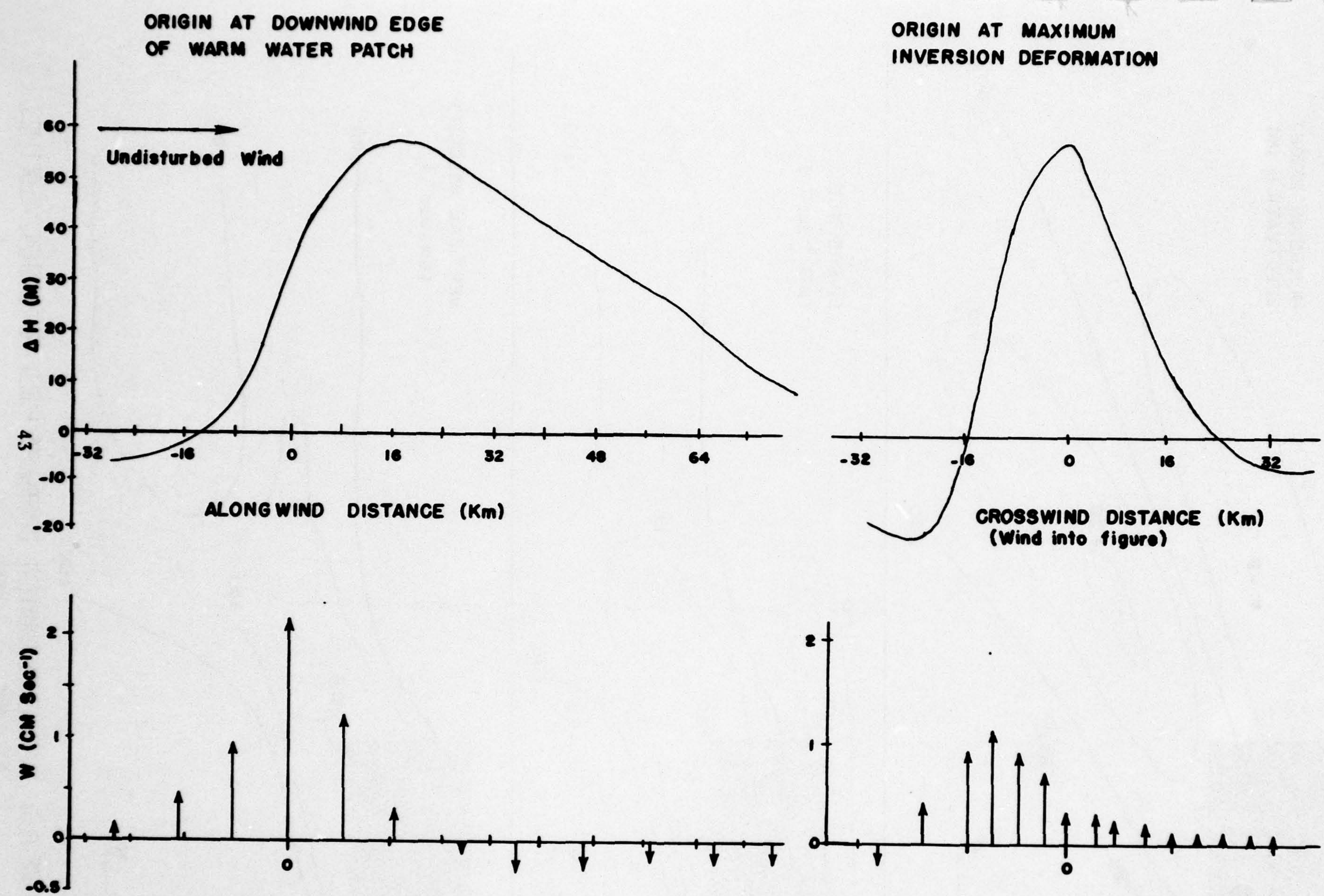


FIGURE A-5: COMPUTED DEFORMATION AND VERTICAL VELOCITY DISTRIBUTION, HI = 300M

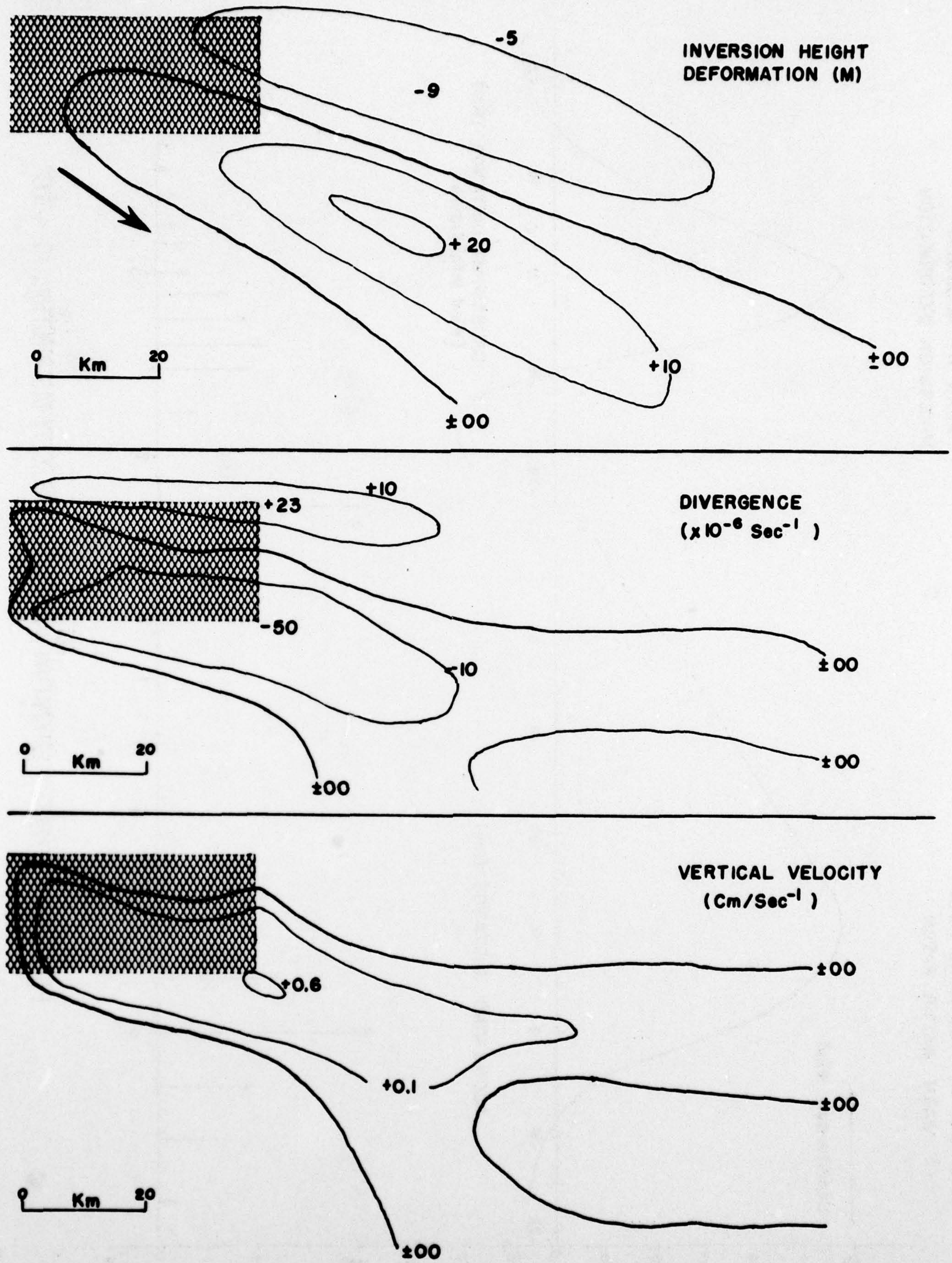
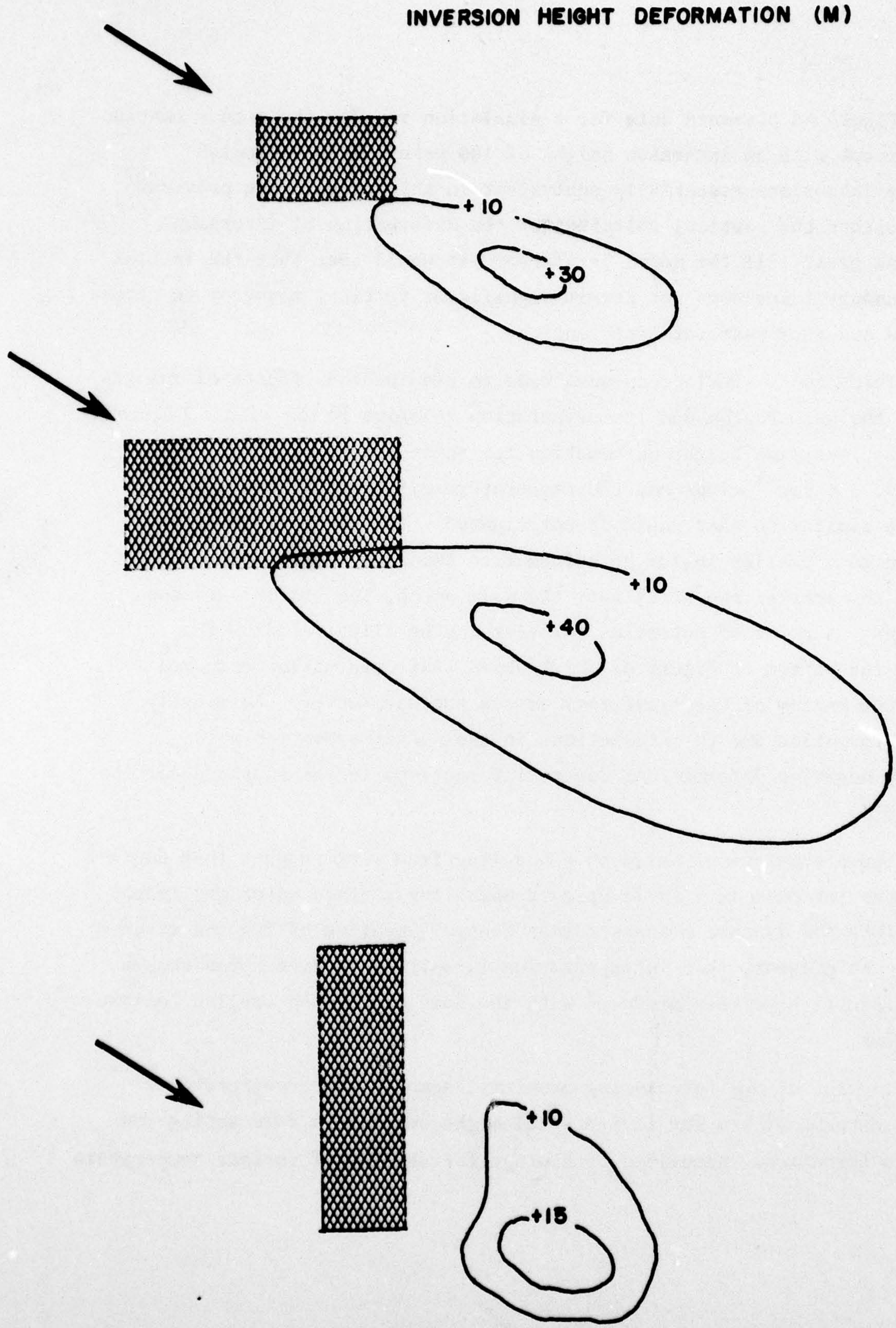


FIGURE A-6: COMPUTED DEFORMATION, DIVERGENCE FIELD AND VERTICAL VELOCITIES, $\Delta T = 2^{0}C$, HI = 100M



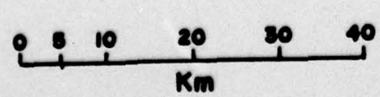


FIGURE A-7: EFFECT OF SIZE AND ORIENTATION ON DEFORMATION OF INVERSION, $\Delta T = 2^{0}C$, HI = 200M

Figure A6 presents data for a simulation run for the same situation as above except with an inversion height of 100 meters. The computed convergence values are essentially equivalent in this case to the previous case but neither the vertical velocity nor the deformation of inversion height is as great. If the model is correct, it would seem that the initial shallow boundary layer does not permit significant vertical momentum to become established and thus restricts deformation.

Three model simulations were made to examine the effects of changes in size of the warm region and its orientation relative to the wind. Figure A7 presents the inversion height deformation for these cases. In each of the three cases, 5 m sec^{-1} winds and $2^{o}C$ temperature differences were used. Results are similar to what would be anticipated. The small patch of warm water produced a smaller region of deformation than the large patch. Similarly, the shorter the fetch over the warm patch, the smaller are the deformations. A point of potential interest may be illustrated by the diagram at the bottom of Figure A7 which shows that deformation occurred only near the bottom of the trajectory across the warm water. Apparently positive deformation due to interactions in the northern portion were canceled by negative deformations due to interactions in the southern portion of this field.

These simulations based on a modeling framework suggest that patches of warm water imbedded in a field of cold upwelling surface water may indeed be responsible for dynamic processes that control location of fog and stratus formation. At present, this interpretation is only conjecture, even though excellent results have been produced with the same model when applied to lake effect storms.

In view of fog forecasting problems, some further conjecture is warranted in terms of how the Lavoie model might be used in forecasting fog and stratus formation. Satellite technology for sensing of surface temperature

is developing rapidly. The major problems are associated with interference effects produced by total precipitable water between the surface and the satellite. Existing instrumentation is capable of detecting relative differences of temperature of the order of 1/2 to 1°C, with spatial resolutions of 1/2 to 1½ nautical miles. Such instrumentation should be capable of providing the surface temperature data (in a relative sense at least) which is required for model input. Wind information for model input can be obtained from large scale weather forecasts. However, two important additional characteristics of the boundary layer must also be included. These are initial inversion height and surface relative humdiity. There may be some potential for determining inversion height from remote temperature measurements of broken stratus tops and adjacent surface water using the assumption that an adiabatic lapse exists beneath the inversion. No ideas have yet been advanced for estimating surface humidity. In any case, we believe that some significant potential exists for developing forecasting techniques based on the Lavoie model.

In the computations used in the foregoing illustrations, the effects of condensation and radiation were not included. Condensation, previously incorporated in the Lavoie model at Calspan for the study of the lake effect storms, was removed for this internal research as an economy measure. It will be reinstated in the immediate future. Radiation has not yet been incorporated into the model, but we are currently investigating procedures for making this addition.

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